

GEOCHEMISTRY OF EASTERN PACIFIC MORB:
IMPLICATIONS FOR MORB PETROGENESIS AND THE NATURE OF CRUSTAL
ACCRETION WITHIN THE NEOVOLCANIC ZONE OF TWO RECENTLY ACTIVE
RIDGE SEGMENTS

By

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Abstract of Dissertation Presented to the Graduate School
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A detailed investigation of two recently volcanically active mid-ocean ridge (MOR) segments in the N.E. Pacific Ocean has resulted in a better understanding of crustal accretion within the neovolcanic zones of medium- to fast-spreading ridges. The use of submersible and other near-bottom systems, in conjunction with rock sampling by wax core, provides excellent spatial control, resulting in a database of precisely located in situ geologic observations and samples. The combination of dense high-resolution sampling, temporal control gained by recent eruptive activity, and multi-year monitoring efforts makes these investigations unique in MOR research.

The CoAxial Segment of the Juan de Fuca Ridge (JdFR) is a medium-spreading rate ridge that has experienced at least three volcanic eruptions between 1981 and 1993. Evidence of amagmatic extension and chemical heterogeneity between different lava flows indicates that the CoAxial Segment has behaved as a magma limited system, despite the occurrence of recent eruptive activity. The current CoAxial Segment neovolcanic zone is greater than 1 km wide, and spatial focussing of eruptive activity is

poor relative to the more magmatically robust Cleft Segment of the JdFR and 9-10° N segment of the East Pacific Rise (EPR). A broader survey of the central JdFR reveals that, on a regional scale, geochemically different melt regimes can be associated with distinct structural, morphologic and volcanic provinces.

Detailed study of ridge-axis morphology, structure, and chemistry of the faster spreading, magmatically robust, 9°-10° N segment of the EPR shows a different relationship between ridge-axis structure and magmatic activity than that observed at the CoAxial Segment. Data from a third-order overlapping spreading center (OSC) located at 9° 37' N suggests that there is a close association between magmatic activity and axial discontinuities, and further suggest that this OSC is a magmatic and hydrothermal divide between adjoining third-order ridge segments. Temporal constraints correlate a recent magmatic event to southward propagation of the eastern OSC limb. These data establish a direct causal relationship between magmatic activity and evolution of ridge-axis discontinuities, and confirm that segmentation of melt delivery to the ridge-axis may affect ridge-axis structure along the fast spreading EPR.

CHAPTER 1 GENERAL INTRODUCTION

The mid-ocean ridge (MOR) system is a globe-encircling volcanic mountain chain more than 60,000 km long. It extends throughout all of the world's major ocean basins, rising some 1000-3000 m above the abyssal sea floor. Mid-ocean ridges are the site of oceanic crustal accretion and lithospheric genesis and, as such, are important to the study of plate tectonics and Earth's geological processes. While an understanding of magmatic processes occurring at MORs is an essential part of understanding plate tectonics, the significance of magmatic activity at ocean ridges extends far beyond geological interests. Magmatic activity along MORs provides the source of heat that drives extensive hydrothermal circulation within the oceanic crust which, in turn, exerts a strong influence on ocean chemistry. Not only does this volcanically driven hydrothermal system help to control seawater chemistry, but it also provides the energy that supports a rich and diverse abyssal ecosystem that utilizes chemosynthetic reduction as the basis of its food web.

Surprisingly, relatively little of this immense volcanic system has been studied in detail. Difficulties in investigating such a remote and extreme environment limited direct examination until the 1960's when advances in technology allowed man to travel to and freely access the mid-ocean ridges, which typically range from 2000-4000 meters below sea level. Early studies sampled the MOR system with coarse spatial resolution, providing insights on some of the first-order differences between different ridges. The

advent of swath mapping systems that allow for rapid underway bathymetric mapping of large areas of seafloor greatly advanced the understanding of MOR morphology and structure.

Discontinuities and offsets in the strike of the ridge-axis separate the MOR system into segments of varying scale, depending on the magnitude of the axial discontinuity. Macdonald et al. (1988) established four different orders of ridge axis discontinuity classified on the basis of the discontinuity's spatial dimensions. In this categorization, which is today generally recognized as the accepted nomenclature, first-order spreading segments are tectonically defined and bounded by large transform offsets or propagating rifts whose offsets are such that plate boundaries behave rigidly, juxtaposing crust of greater than approximately 0.5-1 Myr age difference across the offset boundary. Offsets are typically greater than 50 km and partition the ridge-axis at intervals of 100-1000 km. Second-order segments have a length scale of 50-300 km, and they are bounded by smaller, non-rigid discontinuities such as large overlapping spreading centers (OSC) with an offset greater than 3-5 km or small offset (less than 20 km) non-rigid transform faults. Third-order segmentation divides the ridge-axis into characteristic lengths of 30-100 km, and these segments are bounded by smaller OSCs (0.5 to 3-5 km). These smaller OSCs are often associated with small (10's of meters) increases in axial depths. Lastly, fourth-order segments have typical length scales of 10-50 km and are not commonly associated with increased axial depth near the segment ends. These fourth-order discontinuities are defined by small non-overlapping offsets of the ridge axis and/or small (1° - 5°) bends in the ridge-axis referred to as deviations in axial linearity (DEVALS) [Langmuir et al., 1986]. Although first-order segments and their associated axial discontinuities are

thought to be persistent on the order of millions of years, smaller ridge offsets may not be as temporally stable [Macdonald et al., 1988b].

As investigations of the MOR system were undertaken at finer and finer spatial scales, it became apparent that the mid-ocean ridge environment was the product of complex interactions between magmatic, tectonic, hydrologic, and biologic processes. Only by compiling detailed investigations of processes occurring at different spreading segments would a more comprehensive understanding of this environment be achieved. To date, only a small percentage of the mid-ocean ridge system has been subjected to detailed multi-disciplinary investigation at the first to fourth-order segment scale.

Oceanic ridges exhibit a range of plate divergence or "spreading rates". Full divergence rates vary by over an order of magnitude, and to a first-approximation ridge morphology, structure, and magmatic activity are strongly influenced by this spreading rate [Perfit and Chadwick, 1998]. Typically, slow-spreading ridges (full plate-divergence rate on the order of 1-4 cm/year), exemplified by ridges in the Atlantic and Indian oceans, rise steeply from the abyss relative to faster-spreading MORs. They are generally rifted by an axial graben that can be 10's of kilometers across and greater than 1500 meters in relief. Slow-spreading ridge terrains are generally more greatly affected by extensional tectonics and faulting and, as such, tend to be much rougher than at faster-spreading ridges. Volcanic morphology of the axis floor, which is the site of active crustal accretion, tends to be dominated by discrete constructional features that may coalesce to form mounds or ridges. Lava morphology is often pillowed, a characteristic associated with relatively slow lava effusive rates. The zone of crustal accretion, or "neovolcanic zone," is thought to be relatively wide and unfocussed (2-12 km wide), with relatively

less control on the location of magma emplacement in the crust relative to faster spreading regimes.

Contrary to the above, fast spreading ridge systems ($\sim 8\text{-}16$ cm/yr full-rate), exemplified by the East Pacific Rise, exhibit very different characteristics. Characteristic morphology of fast spreading ridges is that of a broad swell or “rise” in the abyssal seafloor. Fast spreading ridges generally lack a large fault bounded axial valley, and when an axial graben is present, it is of a much smaller scale than those present at slow spreading ridges. Overall ridge structural fabric is less rough, and melt focussing appears to be greater, with a majority of volcanic activity being restricted to a much narrower zone of accretion (100-200 m wide). Magmatic activity appears to be much less episodic than in slower-spreading environments, and overall ridge fabric appears to be more greatly dominated by magmatic processes rather than tectonic processes. Many sections of fast-spreading ridge can be associated with a seismically detectable axial magma chamber in the shallow crust, while these features are largely absent from seismic imaging of slow-spreading ridge-axes. Medium spreading-rate ridges ($\sim 4\text{-}8$ cm/yr full-rate), exemplified by the Juan de Fuca Ridge in the northeast Pacific, exhibit characteristics intermediate between the two end-member axial morphologies (neovolcanic zone 200-2000 m wide).

While recognition of the first-order characteristics has allowed for great advancing in the understanding of mid-ocean ridges, details of the process of oceanic crustal accretion has been hampered by a lack of any high resolution temporal control to the origins of observable phenomenon. The first documented recent eruption was not until the 1990s when anomalies between repeat bathymetric surveys were able to show the

existence of a seafloor volcanic constructional terrain that was not present 6 years prior (time of eruption constrained to being between 1981 and 1987) [Chadwick et al., 1991; Embley et al., 1991]. Since then, fewer than 10 additional eruptions have been confirmed and investigated, several of which are reported on in this dissertation. Numerous questions fundamental to understanding MOR crustal accretion and the linkages between geologic activity and important biologic and oceanographic processes still exist. Outstanding questions, among the as-of-yet unsatisfactorily addressed geologic issues include (but are not limited to): (1) How frequently and with what spatial consistency do volcanic eruptions occur on a given segment of MOR? (2) What is the dependence of eruptive frequency on spreading rate and melt supply? (3) How does spreading rate and melt supply affect melt focussing and the physical configuration of the neovolcanic zone? (4) How does magmatic activity affect ridge axis structure and segmentation or vice-versa? (5) It has been observed that different segments of ridge spreading at similar overall rates can have very different axial morphologies. Is this observation best explained by a temporal cyclicity to the intensity of magmatic and tectonic activity at discrete spreading segments, or do discrete segments have inherent morphologies resulting from differences in volcanic activity that are constant over longer time scales?

Answers to these and other questions that relate to temporal variability and evolution of magmatic, hydrothermal and biological systems can only be answered through the integration of detailed, long-term studies and active monitoring of many different portions of the global MOR system. While this dissertation does not provide holistic answers to any of the above questions, it does present the results of detailed

geologic and petrologic studies of two recently active ridge segments, each occurring in different spreading-rate and apparent melt-supply regimes.

Documentation of recent eruptive activity along two second-order segments of the eastern Pacific mid-ocean spreading system provides a rare opportunity to examine the geometry and petrogenetic associations of magmatic events that form the basis for crustal accretion along medium to fast spreading divergent plate boundaries. The CoAxial Segment of the medium spreading-rate Juan de Fuca Ridge and the 9°-10° N Segment of the fast-spreading East Pacific Rise were the sites of eruptive activity in 1993 and 1991-92 respectively [Fox et al., 1995; Haymon et al., 1993]. The Juan de Fuca Ridge located approximately 500 kilometers west of the Washington/Oregon/British Columbia coast. The 9-10° N latitude segment of the East Pacific Rise is located west of the Central American coast, between the Clipperton and Siqueros fracture zones. Detailed maps of these two regions are provided in chapters 2 and 3 respectively.

Not only are these two second-order spreading segments [Macdonald et al., 1988] different in terms of spreading rate, but they also differ greatly in axial morphology, and possibly current melt supply rates from the upper mantle. The 9°-10° N segment of the East Pacific Rise has an axial morphology typical of faster spreading ridge segments dominated by magmatic activity. Contrary to this, the CoAxial segment has an axial morphology intermediate in nature between the fast- and slow-spreading end-members, and is much more indicative that amagmatic extension plays a prominent role in the development of axial morphology. Additionally, these two recently active ridge segments provide a good contrast to the Cleft Segment, another comparably studied portion of the Juan de Fuca Ridge that although spreading at a similar rate to the CoAxial

segment, displays many of the volcanic features observed on the EPR [Smith et al., 1994]. The two investigations presented in this dissertation, and the results that arose from them, are briefly summarized below in “Abstract form” for the convenience of the reader, providing a bit more detail than is presented in the overall dissertation abstract.

1.1. The CoAxial Segment

Recent investigations along the central portion of the Juan de Fuca Ridge have identified at least three different eruptive units that were emplaced along the CoAxial Segment between 1981 and 1993 [Chadwick et al., 1995]. Major element, trace element and isotopic compositions of mid-ocean ridge basalt (MORB) samples recovered by submersible, ROV and wax core are examined in order to critically assess the spatial and temporal chemical variability on regional and local scales. The data allow us to quantify the chemical variances within and between each of the mapped flow units, identify and define genetic relationships between the flows, and to document the relationship of these recent eruptives to older basalts recovered along the CoAxial and other nearby segments of the Juan de Fuca Ridge.

Geochemical data show that each of the three recent eruptives are distinct from one another, supporting the hypothesis that at least three separate volcanic units have been emplaced along the CoAxial segment since 1981. Further, when compared to fresh looking eruptives sampled from within Axial Seamount caldera and along its associated north and south rift zones, recent CoAxial eruptives have isotopic and elemental characteristics distinct from lavas associated with Axial Seamount volcanism. This suggests that magmas parental to lavas sampled from each of these regions have unique origins, and that it is highly unlikely that Axial Seamount was the source of the 1993

CoAxial eruption, as has been suggested. CoAxial segment lavas are geochemically very depleted and show no geochemical evidence for influence of Axial Seamount hotspot.

The integration of these geochemical data with observational, photographic and sidescan sonar data defines several different tectono-magmatic provinces in the region between the Vance and Cobb segments of the Juan de Fuca Ridge. The presence of these different provinces suggests quite distinct sub-axial melting regimes exist along this medium rate spreading ridge.

1.2. The East Pacific Rise between 9° and 10°N

Alvin based investigations of a small ridge-axis discontinuity, located at the crest of the East Pacific Rise crest between ~9° 35-37'N, provide strong evidence that the feature is actually a small overlapping spreading center (OSC) that divides two magmatic and hydrothermal systems. Two overlapping axial summit collapse troughs (ASCT) morphologically express this third-order ridge-axis discontinuity. The eastern ASCT that extends from the north has an approximately 0.4 km right-lateral offset from the western ASCT that extends from the south and overlaps it by at least 1.2 km. Changes in volcanogenic morphology and increased hydrothermal and biologic activity observed between surveys of the area in 1989 and 1991 suggest that a magmatic event affected this region during this time-period. These changes combined with geochemical data from basalt samples and hydrothermal fluids recovered during three *Alvin* dives suggest that the eastern limb of the OSC is actively propagating to the south. Observational data that support this hypothesis include: (1) the occurrence in 1991 of extensive areas of bacterial bloom and diffuse venting of 39°C, low-chlorinity hydrothermal fluids along the southern extension of the eastern ASCT; (2) regions of local collapse of drained out young-

looking lava flows, the presence of lava pillars, and hydrothermal fluids venting directly out of glassy, young-looking lava flows along the southward extension of the trend of the eastern ASCT; and (3) the presence of several extinct sulfide edifices within the western ASCT, and venting of high chlorinity hydrothermal fluids at temperatures of greater than 113°C in 1991 and 1994 within that region.

MORB geochemistry lends further support to the hypothesis that southward propagation of the eastern OSC limb at 9° 37' N is related to recent magmatic activity in the 9° 37-52' N area. Basalt samples with the highest MgO (most primitive, highest temperature) are typically the freshest-looking lavas and tend to be spatially associated with the ASCT of the eastern OSC limb, and its extension to the south. More fractionated (lower temperature) basalt samples are found along the western limb ASCT and in the zone of overlap between the two summit troughs. The more evolved MORB western ASCT MORB are consistently older in appearance than the fresher, young-looking high-MgO lavas to the east. Recently enhanced magmatic activity of the OSC's eastern limb is apparent by MgO content of the most recent eastern ASCT lavas which have Mg#'s that are anomalously high relative to the trend observed in the regional MORB geochemical data. These lavas have major element chemistry similar to the 1991 eruptive at 9° 46-52' N and may be related to the same phase of magmatic activity. Additionally, the chemistry of hydrothermal fluids recovered both within the region of the OSC and along segments of the EPR crest proximal to it, indicates that the OSC marks a hydrographic divide. Hydrothermal fluids generally have higher chlorinity than seawater south of the OSC, and a lower chlorinity than seawater north of it (including fluids from the eastern OSC limb). Investigation of hydrothermal fluids between 9° 17-

54°N have shown that low-chlorinity fluids are associated with either eruptive/intrusive events, or occur in areas with very shallow and robust heat sources. With time, the fluids often increase in chlorinity. It is only at the 9°37'N OSC and further south that extensive extinct sulfides and fluids with chlorinities ~1.5 times seawater have been sampled, suggesting that hydrothermal systems north of the OSC have been more recently perturbed by a magmatic event than those on the failing rift and further south. It is proposed that the OSC's southward propagation is locally driven, resulting from a combination of renewed and enhanced magmatic activity north of the propagator, and a waning magmatic system south of it. Local occurrence of E-MORB on the Pacific plate near the OSC may reflect lower degrees of melting in the region of the retreating OSC limb.

CHAPTER 2
GEOCHEMICAL CHARACTERISTICS AND PETROGENETIC RELATIONSHIPS
OF RECENTLY ERUPTED MORB FROM THE CENTRAL JUAN DE FUCA RIDGE:
SEGMENT SCALE, INTERFLOW, AND INTRAFLOW VARIATIONS

2.1. Introduction

2.1.1 Field Investigations and Previous Studies

The 1993 detection of a seismic swarm thought to be related to dike intrusion along what is now known as the CoAxial Segment of the Juan de Fuca Ridge [Fox et al., 1995] triggered an extensive series of investigative cruises to this portion of the mid-ocean ridge system. A total of three rapid response cruises undertaken between July and October of 1993 (the seismic event lasted from June 26 until July 10, hydrocasting began on July 3 and the first direct sea floor observations began on July 9) investigated the remotely detected seismic events. Investigators participating on these cruises sampled hydrothermal event plumes and visually confirmed that a basalt flow had indeed erupted onto the ocean floor in the location where the most intense seismic swarms had been located [Embley et al., 1995; Baker et al., 1995]. These cruises concentrated on mapping and sampling lavas, biota, and hydrothermal waters associated with the new flow and along the path of proposed dike emplacement. Subsequently, at least 7 cruises to this region have expanded the sampling and mapping efforts to include other portions of the CoAxial Segment as well as Axial Seamount and its associated north and south rift zones.

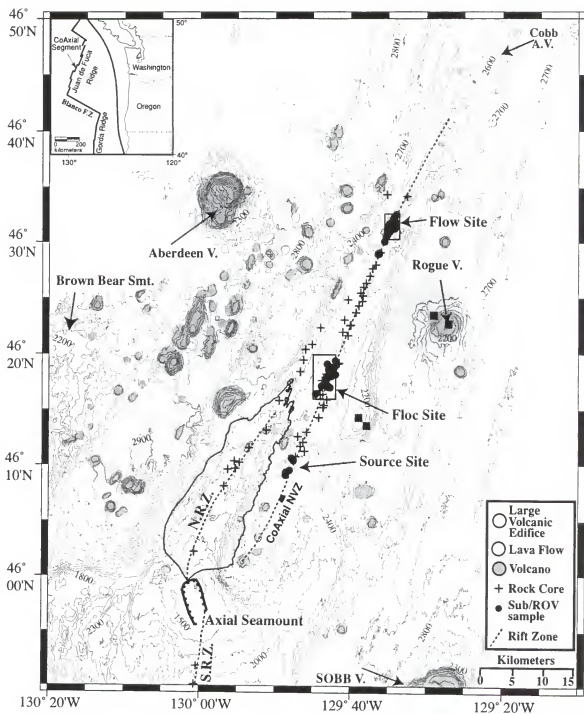
Prior to the aforementioned events, very little was known from this section of the Juan de Fuca Ridge (with the exception of Axial caldera). The CoAxial Segment had not been recognized as a distinct second-order MOR segment [Macdonald and Fox, 1983; Embley et al., 1995], and geochemical data from only four relatively poorly located dredges had been reported in the literature (Eaby et al., 1984; Rhodes et al., 1990).

2.1.2 Regional Geology

The geology and morphology of the CoAxial Segment and its surrounding terrain is discussed in detail by Embley et al. [in press], and, as such, will only be discussed here in sufficient detail to provide basic background information. This paper is, in many ways, complimentary to and an extension of a recent paper by Embley et al. [in press].

The Juan de Fuca Ridge (JdFR) is a first-order, medium spreading rate (~ 6 cm/yr full rate) ridge segment located approximately 500 km off of the Oregon/Washington/ British Columbia coast (Figure 2-1). It extends some 490 km along strike ($\sim 020^\circ$) between the Blanco and Sovanco Fracture Zones which intersect the ridge-axis at $\sim 44^\circ 30'$ and $48^\circ 45'$ N latitude respectively. The JdFR has been subdivided by various authors into at least 6 different second-order ridge segments ranging from 50–150 km in length. More detailed discussions of the JdFR and its second-order subdivisions have been published [Delaney et al., 1981; Johnson and Holmes, 1989; Kappel and Ryan, 1986; Kappel and Normark, 1987; Hey, 1982]. Axial Seamount, a large ridge-axis centered volcano that is the latest expression of the Cobb melting anomaly [Desonie and Duncan, 1990], is located in the middle portion of the JdFR at $\sim 45^\circ 55'$ N latitude and 130° W longitude (Figure 2-1). This large edifice rises more than 1000 m from the surrounding abyssal terrain and has produced a large summit caldera and two major rift

Figure 2-1: Bathymetric map of the CoAxial Seamount region taken after Figure 4 of Embley et al., in press. True north is towards the top of the page. Regional NE Pacific map is shown in inset (upper left). Shading depicts interpretation of volcanic terrain based on sidescan sonar and bathymetric data (Embley et al., in press). Basalt sample locations are shown as crosses (wax core samples), filled circles (ROV/Submersible samples) and filled squares (starting/ending points of dredges). Solid black outline depicts the interpreted aerial extent of Axial Seamount north rift zone volcanism (see text). Dashed lines mark major rift zones of Axial Seamount and the CoAxial Segments. Rectangular outlines mark the Floc and Flow Sites which are shown in Figures 2-2a (southern outline) and 2-2b (northern outline) respectively. Regional shading is based on interpretation of sidescan and bathymetric data and is keyed to the legend (lower right).



zones extending north (to $\sim 46^{\circ}15' \text{ N}$) and south (to at least $45^{\circ}30' \text{ N}$) from the main edifice, which is some 20 km in diameter [Johnson and Holmes, 1989]. Though this edifice is the youngest volcano of the Cobb-Eickelberg chain, a greater than 300 km long seamount chain, Axial Seamount lavas do not display geochemical enrichments typically associated with hot spot volcanism [Rhodes et al., 1990]. Together, Axial Seamount and its north and south rift zones comprise the Axial Seamount Province.

The southern end of the Coaxial Segment exists in an overlapping relationship with the Axial Seamount's north rift zone (NRZ) to its west, overlapping it by at least 25 km. The southern terminus of the CoAxial segment is at Helium Basin ($\sim 46^{\circ} \text{ N}$ and $129^{\circ}55' \text{ W}$) which lies at the base of the steep, faulted NE flank of Axial Seamount [Lupton, 1990]. The CoAxial Segment extends north-northeast along a strike of $\sim 022^{\circ}$ to $\sim 46^{\circ}40'$ where it overlaps some 10 km with the Cobb segment to its east [Embley et al., in press]. The ridge is shallowest at its southern end, and deepens northward along the strike of the ridge by approximately 425 m to a depth of $\sim 2550 \text{ m}$ at its northern terminus. The CoAxial Segment's axial valley is quite wide compared with other JdFR second-order segments (up to 2 km), as is the apparent current zone of active crustal accretion, which will be discussed further below. It is intriguing that the only other segment on the JdFR with as wide of an axial valley is the Vance Segment which lies immediately south and with Axial Volcano's south rift zone (SRZ) [Embley et al., in press].

Based upon our sampling regimen and the criteria presented in Embley et al. [in press], we have designated three different sites along the CoAxial segment as the primary focus of our investigations. These sites listed from south to north are the "Source Site",

the “Floc Site” and the “Flow Site” (Figure 2-1). Each of these sites is described in detail in Embley et al. [in press], and the reader is referred to that paper for a detailed discussion of their geology and morphology.

2.2. Data Collection and Sample Distribution

2.2.1 Alvin and ROV Based Sample Acquisition

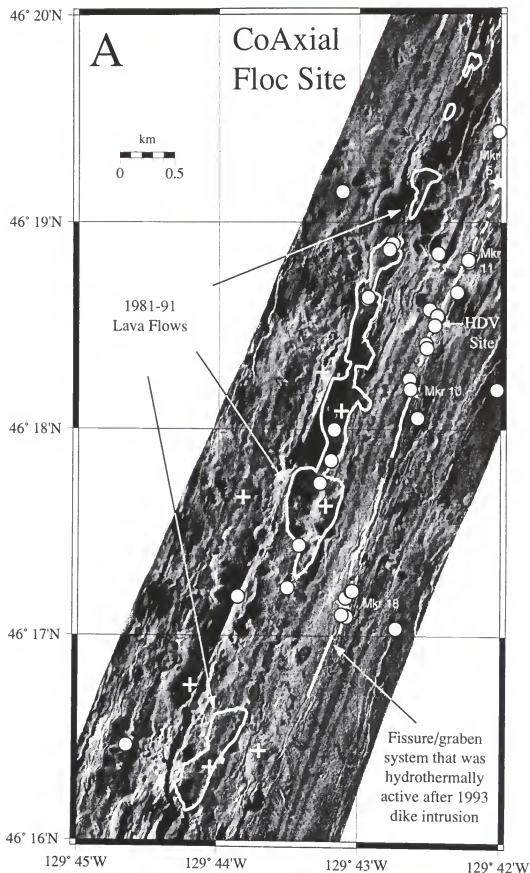
A total of 136 site-specific *in situ* basalt samples have been recovered from the CoAxial segment since the first rapid response cruises in late 1993 (Figures 2-1 and 2-2). These samples were collected utilizing a variety of submersible (104 samples) and remotely operated vehicle (ROV) platforms (32 samples) employed during 7 different oceanographic expeditions conducted between 1993 and 1996.

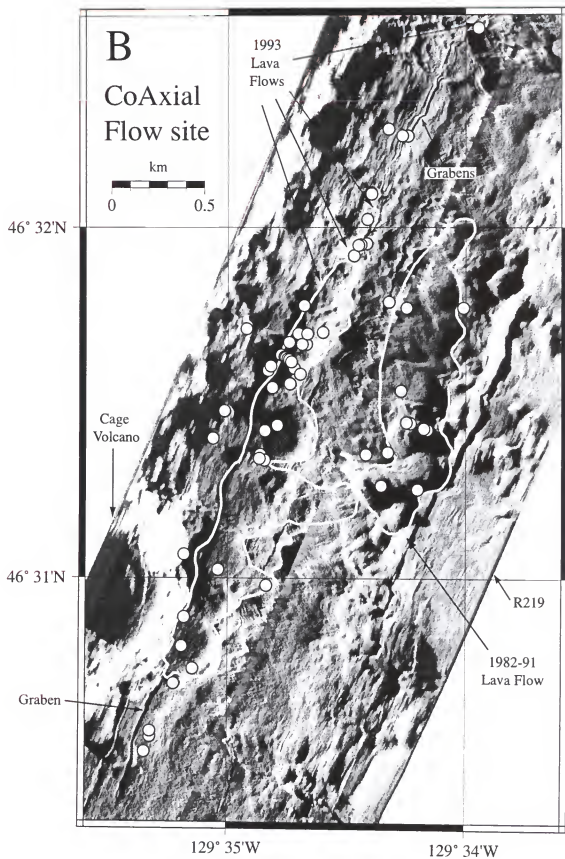
The majority of ROV and submersible sampling concentrated around the three primary study sites (described above and in Embley et al. [in press]), with an emphasis placed on sampling lavas of different apparent age and flow morphology. In addition, special attention was paid to recovering a denser and well distributed suite of samples for areas where discrete eruptive units could be identified and mapped on the basis of visual observations and depth differences between repeat bathymetric surveys [Chadwick et al., 1995]. Relative ages and flow contact relationships were deduced from stratigraphic relationships, flow morphology, amount of sediment cover and presence or absence of sessile organisms.

2.2.2 Surface Based Sample Acquisition

The suite of basalt samples recovered by submersible and ROV is supplemented by 57 samples recovered by wax corer (out of 66 attempts), two dredge hauls and one TV Grab (Figures 2-1 and 2-2). Surface based sampling was generally performed at night or

Figure 2-2: Expanded view of the Floc (A) and Flow (B) sites depicted by rectangle outlines in Figure 2-1. Figures are after figures 11 and 8 of Embley et al., in press. Basalt sample locations, mapped boundaries of recent flows discussed in text, major fissures and low temperature hydrothermal venting locations are superimposed on images generated by deep tow AMS 60 sidescan sonar (Embley et al., in press). Flow boundaries were determined in consort with visual and photographic mapping, repeat bathymetric surveying and sidescan data. Wax core samples are shown as crosses, and submersible/ROV collected samples as circles. Fissures marked by bold white lines (2-2A) depict the current zone of hydrothermal activity, and interpreted ridge-axis, which extends northward passing directly underneath the 1993 Flow (2-2B).





during submersible/ROV down time. Wax cores primarily sampled the Axial Seamount NRZ (15) and regions along the CoAxial segment outside of the major sites of submersible investigation. Dredges sampled Rogue Volcano and a topographic high bounding the east side of the CoAxial Segment. The video guided grab sample recovered basalt from a fresh flow outcropping along Axial Seamount NRZ. This flow was also sampled by wax core.

2.2.3 Sidescan Sonar and Near Bottom Photographic Data Acquisition

Photographic data gathered by submersible, ROV and near bottom camera tow, as well as sidescan imagery acquired using SeaMARC II and AMS-60 sidescan sonar systems, have been incorporated in our regional geologic interpretations. Submersible, ROV and camera tow data were generally located using long-baseline transponder navigation that provides positional accuracy to within approximately 20 meters. AMS-60 sidescan data have a spatial resolution of 1m. Details of the distribution of both the photographic data and the sidescan sonar imaging used to place samples in a geologic context are presented in Embley et al. [in press].

2.3. Analytical Methods

2.3.1 Sample Preparation

All basalt samples were described and catalogued at sea. Geologic context and the location for each sample were carefully documented, and glassy rinds sub-sampled when present. Shoreside, a representative suite of samples were sectioned for petrographic analyses. Natural basaltic whole rock glasses (quenched glass $\pm$ crystals) were cleaned and prepared for analysis by x-ray fluorescence. Glasses were rinsed in 2X distilled water and the cleanest portion (generally 5-50 grams for *ALVIN*/ROV recovered

samples) of the glass was hand picked under binocular microscope for further preparation. Picked glass subsamples were subsequently coarsely crushed and sonicated first in a solution comprised of equal parts 2.5N HCL and 30% H₂O₂ (15-30 min) and then in 2X distilled water. Samples were then dried and picked free of any visible surficial alteration or precipitated coating using a binocular microscope. Glass splits to be analyzed by electron microprobe, thermal ionization mass spectrometry (TIMS) or laser ablation ICP-MS (LA ICP-MS) were then further hand picked to assure that only the most crystal-free and pristine glass was analyzed. Glass splits to be analyzed by XRF or solution ICP-MS were ground using either a tungsten-carbide disk mill, a SPEX agate ball mill, or some combination of the two. Use of the tungsten-carbide disk mill was kept to a minimum, but it is likely that in at least some of the samples the elements Ta and W may be to some degree compromised by tungsten-carbide contamination [Thompson and Bankston, 1970]. Samples (< 100 mg) analyzed by TIMS were subjected to a second leaching (~10 minutes) using warmed distilled 6N HCL and digested in a solution of distilled HF and ultra pure HNO₃. Dissolved glasses were then chlorinated using distilled 6N HCL and Sr and Nd separated and collected using standard ion chromatographic techniques.

2.3.2 Major Element Analyses

The major element abundances of 153 natural basaltic glasses are reported here and a representative subset of the data are presented in Tables 2-1 and 2-2. The majority of these analyses were measured using a JEOL 8800 electron microprobe (EMP) at the USGS in Denver. Multiple standards were frequently measured during the course of each analytical run, including the mid-ocean ridge basalt (MORB) glass standards JdF-D2 and

Table 2-1. Electron microprobe Analyses of Representative Natural Glasses Recovered by Submersible/ROV

Sample ID	Latitude N.	Longitude W.	Loc.	#	SiO ₂	Al ₂ O ₃	TiO ₂	FeO	MnO	MgO	CaO	Na ₂ O	K ₂ O	P ₂ O ₅	Total
221-0445	46° 30.743'	129° 35.147'	1	10	50.6	13.8	1.61	12.5	0.23	6.90	11.4	2.65	0.12	0.14	99.9
2670-4	46° 31.553'	129° 34.739'	1	8	50.4	13.5	1.69	12.9	0.24	6.69	11.4	2.66	0.12	0.14	99.6
2672-2	46° 31.353'	129° 34.864'	1	10	50.5	13.6	1.67	12.7	0.24	6.77	11.4	2.67	0.12	0.14	99.7
2788-4R	46° 31.598'	129° 34.721'	1	9	50.5	13.6	1.66	12.7	0.22	6.73	11.3	2.65	0.13	0.15	99.6
2792-1R-2	46° 31.778'	129° 34.678'	1	6	50.5	13.5	1.70	12.9	0.23	6.65	11.3	2.71	0.12	0.17	99.6
2794-2R	-46° 32.56'	-129° 33.974'	1	10	50.6	13.5	1.65	12.7	0.23	6.67	11.3	2.65	0.12	0.14	99.6
2983-R2	46° 31.636'	129° 34.775'	1	9	50.5	13.6	1.68	12.8	0.24	6.72	11.3	2.63	0.13	0.15	99.8
2983-R6	46° 31.624'	129° 34.749'	1	10	50.5	13.5	1.70	12.9	0.24	6.70	11.3	2.69	0.13	0.15	99.8
2993-R4	46° 31.024'	129° 35.040'	1	5	50.3	13.6	1.67	12.6	0.25	6.79	11.5	2.59	0.11	0.14	99.5
JA187-7	46° 31.696'	129° 34.665'	1	10	50.5	13.6	1.68	12.8	0.23	6.73	11.2	2.65	0.12	0.15	99.7
2672-6	46° 31.266'	129° 34.353'	2	8	49.4	14.3	2.08	12.7	0.22	6.94	10.9	2.71	0.15	0.20	99.5
2792-4R	46° 31.788'	129° 34.321'	2	7	49.6	14.4	2.02	12.3	0.20	6.95	10.8	2.74	0.16	0.21	99.4
2792-5R	46° 31.772'	129° 34.247'	2	3	49.6	14.4	2.03	12.4	0.23	6.83	10.8	2.77	0.17	0.19	99.4
2792-7R	46° 31.534'	129° 34.270'	2	9	49.3	14.8	1.93	12.3	0.22	7.20	10.9	2.77	0.15	0.19	99.7
2985-R1	46° 31.424'	129° 34.161'	2	10	49.4	14.6	1.97	12.4	0.23	7.11	10.8	2.75	0.16	0.20	99.5
2985-R3	46° 31.442'	129° 34.232'	2	10	49.4	14.6	1.97	12.3	0.22	7.06	10.9	2.76	0.16	0.20	99.6
JA187-4	46° 31.475'	129° 35.014'	2	9	49.6	14.7	1.95	12.1	0.22	7.06	10.9	2.76	0.17	0.19	99.6
2948-8R	46° 17.439'	129° 43.425'	3	7	50.5	14.3	1.45	11.1	0.21	7.27	11.9	2.62	0.15	0.16	99.6
2948-9R	46° 17.742'	129° 43.276'	3	10	50.3	14.4	1.40	10.9	0.21	7.50	12.0	2.6	0.14	0.13	99.6
2948-11R	46° 17.998'	129° 43.177'	3	8	50.4	14.4	1.40	11.1	0.20	7.39	11.9	2.68	0.15	0.16	99.8
2995-R2	46° 18.610'	129° 42.940'	3	5	50.4	14.3	1.55	11.3	0.18	7.21	11.7	2.58	0.14	0.14	99.5
JA188-1a	46° 18.632'	129° 42.944'	3	5	50.4	14.3	1.48	11.4	0.21	7.17	11.8	2.74	0.16	0.15	99.7
JA188-3	46° 18.898'	129° 42.761'	3	8	50.6	14.0	1.50	11.5	0.22	7.01	11.8	2.69	0.16	0.16	99.7
JA188-5	46° 18.898'	129° 42.761'	3	9	50.6	14.1	1.52	11.6	0.20	7.03	11.7	2.69	0.16	0.17	99.7
2791-1R	46° 18.416'	129° 42.525'	4	5	50.2	14.4	1.43	10.9	0.16	7.94	12.3	2.48	0.08	0.11	100.0
2791-5R	46° 18.522'	129° 42.461'	4	5	50.0	14.4	1.46	10.8	0.15	7.85	12.2	2.56	0.08	0.11	99.7
2793-5R	46° 18.241'	129° 42.642'	4	5	50.2	14.5	1.29	10.9	0.20	7.93	12.4	2.48	0.08	0.11	100.1
2946-1R	46° 18.850'	129° 42.440'	4	5	50.1	14.3	1.45	10.7	0.21	7.83	12.4	2.48	0.08	0.11	99.7
2946-2R	46° 18.819'	129° 42.228'	4	5	50.2	14.5	1.47	10.8	0.18	7.69	12.3	2.47	0.09	0.11	99.9
2947-1R-2	46° 18.549'	129° 42.446'	4	5	50.2	14.6	1.41	10.6	0.24	7.91	12.4	2.47	0.08	0.12	100.0
2948-10R	46° 17.850'	129° 43.199'	4	2	50.6	14.5	1.35	10.5	0.24	7.76	12.3	2.44	0.09	0.13	99.7
2948-1R	46° 17.035'	129° 42.737'	4	4	50.6	14.5	1.28	10.2	0.16	7.73	12.6	2.08	0.10	0.12	99.3
2948-2R	46° 17.092'	129° 43.087'	4	2	50.6	14.2	1.30	10.7	0.18	7.85	12.2	2.41	0.08	0.14	99.6

Sample Location Key: 1-1993 Flow Site Lava; 2-1980s Flow Site Lava; 3-1980s Flow Site Lava; 4-Flow Site; 5-Flow Site; 6-Source Site; 7-Other CoAxialSegment; 8- Western Fault Block Ridge; 9-Axial Seamount North Rift Zone; 9*-Recent Flow in NRZ; 0-Small Volcanoes in NRZ; 11-Rogue Volcano.

Table 2-1 continued.

Sample ID	Latitude	N. Longitude	W. Loc.	#	SiO ₂	Al ₂ O ₃	TiO ₂	FeO	MnO	MgO	CaO	Na ₂ O	K ₂ O	P ₂ O ₅	Total
2948-6R	46° 17.185'	129° 43.092'	4	2	49.4	14.7	1.33	10.7	0.27	8.29	12.7	2.23	0.12	0.12	99.8
2948-7R	46° 17.232'	129° 43.507'	4	3	50.4	14.3	1.29	10.8	0.19	7.76	12.2	2.48	0.08	0.13	99.7
2949-1R	46° 17.102'	129° 43.121'	4	5	50.3	14.3	1.31	10.7	0.23	7.89	12.3	2.43	0.07	0.09	99.6
2949-4R	46° 17.218'	129° 43.053'	4	5	50.5	14.4	1.31	10.6	0.21	7.94	12.0	2.42	0.07	0.10	99.6
2949-5R	46° 20.599'	129° 33.525'	4	6	50.3	14.5	1.30	10.6	0.20	7.86	12.3	2.37	0.11	0.12	99.7
221-0551	46° 30.807'	129° 35.194'	5	5	48.4	17.3	1.04	9.0	0.17	9.29	12.1	2.43	0.04	0.06	99.8
2788-6R	46° 31.591'	129° 34.831'	5	4	49.4	15.5	1.19	9.8	0.24	8.25	12.5	2.62	0.06	0.08	99.6
2792-1R-1	46° 31.778'	129° 34.677'	5	10	48.9	16.2	1.15	9.6	0.18	8.63	12.2	2.6	0.06	0.08	99.7
2792-2R	46° 31.955'	129° 34.617'	5	3	49.8	14.4	1.98	12.2	0.23	6.87	10.9	2.77	0.17	0.20	99.6
2792-8R	46° 31.354'	129° 34.418'	5	5	50.0	13.7	1.76	12.2	0.24	7.01	11.8	2.82	0.09	0.14	99.7
2794-1R	46° 32.024'	129° 34.412'	5	5	50.0	13.6	1.63	12.3	0.28	6.99	11.5	2.59	0.09	0.13	99.1
2983-R1	46° 31.711'	129° 34.921'	5	5	48.9	15.5	1.52	10.7	0.22	7.68	11.4	2.83	0.11	0.15	99.0
2983-R5	46° 31.624'	129° 34.749'	5	5	48.1	17.1	1.04	8.8	0.20	9.56	12.1	2.34	0.03	0.07	99.3
2989-R1	46° 31.951'	129° 34.440'	5	5	49.0	16.0	1.19	9.6	0.20	8.62	12.2	2.55	0.04	0.07	99.5
2989-R5	46° 31.772'	129° 34.009'	5	5	48.9	14.1	2.15	12.9	0.22	6.70	10.9	2.75	0.15	0.21	99.0
2989-R6	46° 32.096'	129° 34.395'	5	5	50.1	13.7	1.66	12.1	0.23	6.88	11.5	2.66	0.08	0.13	99.1
2990-R1	46° 28.994'	129° 36.409'	5	5	49.7	14.3	1.48	10.8	0.21	7.70	12.4	2.61	0.07	0.12	99.5
2990-R2	46° 29.071'	129° 36.279'	5	5	50.3	14.8	1.18	9.9	0.22	8.06	12.7	2.52	0.05	0.10	99.8
2990-R3	46° 28.984'	129° 36.281'	5	5	50.0	14.3	1.53	10.8	0.24	7.49	12.5	2.61	0.07	0.12	99.5
2991-R3	46° 32.280'	129° 34.322'	5	5	49.5	15.6	1.18	9.7	0.22	8.39	12.6	2.52	0.05	0.07	99.8
2991-R4	46° 32.280'	129° 34.322'	5	5	49.3	15.9	1.19	9.7	0.19	8.53	12.3	2.53	0.04	0.08	99.8
2993-R2	46° 30.980'	129° 34.838'	5	5	50.4	13.7	1.66	12.5	0.24	6.81	11.4	2.63	0.11	0.15	99.7
2993-R5	46° 31.068'	129° 35.185'	5	5	48.3	17.2	1.06	9.1	0.15	9.39	12.0	2.41	0.03	0.06	99.7
2787-1R	46° 09.404'	129° 48.524'	6	5	50.3	13.8	1.63	12.1	0.23	6.97	11.7	2.59	0.12	0.15	99.6
2790-1R	46° 09.137'	129° 48.615'	6	5	50.4	14.1	1.44	11.4	0.19	7.53	11.9	2.51	0.11	0.13	99.7
2790-4R	46° 09.336'	129° 48.589'	6	5	50.4	14.0	1.49	11.4	0.18	7.41	11.8	2.54	0.11	0.13	99.5
2945-5R	46° 09.342'	129° 48.588'	6	5	50.5	14.2	1.48	11.1	0.23	7.38	12.0	2.51	0.13	0.15	99.6
1993 Flow Site Unit	Average(n=24)				50.5	13.6	1.68	12.8	0.24	6.73	11.3	2.67	0.13	0.15	
	±2 Std. Dev.				0.2	0.2	0.05	0.2	0.02	0.11	0.1	0.06	0.01	0.02	
1980s Flow Site Unit	Average(n=10)				49.4	14.5	2.00	12.4	0.23	7.03	10.9	2.75	0.16	0.20	
	±2 Std. Dev.				0.4	0.3	0.12	0.4	0.02	0.22	0.1	0.04	0.01	0.02	
1980s Floc Site Unit	Average(n=12)				50.5	14.2	1.45	11.2	0.20	7.28	11.9	2.63	0.15	0.15	
	±2 Std. Dev.				0.2	0.3	0.21	0.6	0.03	0.55	0.5	0.19	0.03	0.05	

Sample Location Key: 1-1993 Flow Site Lava; 2-1980s Flow Site Lava; 3-1980s Floc Site Lava; 4-Floc Site; 5-Flow Site; 6-Source Site; 7-Other CoAxialSegment; 8- Western Fault Block Ridge; 9-Axial Seamount North Rift Zone; 9*-Recent Flow in NRZ; 0-Small Volcanoes in NRZ; 1-1-Rogue Volcano.

Table 2-2. Electron microprobe Analyses of Natural Glasses Recovered by Rock Core, Dredge or Video Grab

Sample ID	Latitude N.	Longitude W.	Loc.	#	SiO ₂	Al ₂ O ₃	TiO ₂	FeO	MnO	MgO	CaO	Na ₂ O	K ₂ O	P ₂ O ₅	Total
94RC19	46° 16.358'	129° 44.052'	3	9	50.5	14.5	1.16	10.3	0.19	8.0	12.6	2.37	0.11	0.09	99.7
94RC20	46° 18.089'	129° 43.127'	3	8	50.6	14.2	1.46	11.3	0.22	7.1	11.8	2.69	0.16	0.15	99.6
95 RC43	46° 10.790'	129° 47.280'	7	2	50.0	14.6	1.45	10.6	0.22	7.6	12.2	2.85	0.10	0.12	99.8
95 RC42a	46° 11.279'	129° 46.107'	7	2	49.6	15.4	1.23	10.1	0.15	8.2	12.2	2.55	0.08	0.12	99.7
95 RC42b	46° 11.279'	129° 46.107'	7	3	50.2	13.8	1.86	12.4	0.20	6.5	11.2	3.21	0.18	0.17	99.7
95 RC41	46° 11.707'	129° 46.650'	7	5	50.2	13.9	1.85	12.2	0.21	6.5	11.2	3.08	0.18	0.17	99.5
95 RC40	46° 12.089'	129° 46.318'	7	4	50.4	13.8	1.69	12.5	0.19	6.5	11.3	2.94	0.14	0.15	99.7
95 RC39	46° 12.595'	129° 46.987'	7	4	50.4	13.9	1.67	12.1	0.21	6.8	11.4	2.86	0.15	0.15	99.6
95 RC38	46° 12.989'	129° 45.825'	7	5	50.4	13.8	1.79	12.3	0.23	6.7	11.3	2.93	0.16	0.18	99.7
95 RC13	46° 14.308'	129° 44.231'	7	5	50.9	14.3	1.30	10.4	0.20	7.8	12.3	2.41	0.09	0.10	99.7
95 RC14	46° 15.238'	129° 43.684'	7	5	50.7	14.0	1.37	11.1	0.21	7.5	11.9	2.54	0.09	0.13	99.6
95 RC16	46° 15.431'	129° 43.540'	7	5	50.4	14.6	1.23	10.3	0.21	8.2	12.4	2.40	0.09	0.14	99.9
95 RC15	46° 15.950'	129° 43.163'	7	5	50.5	14.7	1.32	10.5	0.20	7.8	12.2	2.42	0.14	0.13	99.8
94 RC19	46° 16.358'	129° 44.052'	7	9	50.5	14.5	1.16	10.3	0.19	8.0	12.6	2.37	0.11	0.09	99.7
94 RC18	46° 16.441'	129° 43.703'	7	5	50.5	14.4	1.32	10.4	0.21	7.8	12.3	2.48	0.11	0.13	99.6
94 RC17	46° 16.760'	129° 44.197'	7	5	50.5	14.1	1.47	11.3	0.22	7.4	11.9	2.63	0.11	0.13	99.8
94 RC15	46° 17.631'	129° 43.236'	7	9	50.6	14.1	1.49	11.5	0.22	7.0	11.7	2.72	0.15	0.15	99.7
94 RC16	46° 17.674'	129° 43.824'	7	5	50.7	14.0	1.47	11.4	0.22	7.3	11.9	2.62	0.12	0.14	99.8
94 RC20	46° 18.090'	129° 43.127'	7	8	50.6	14.2	1.46	11.3	0.22	7.1	11.8	2.69	0.16	0.15	99.6
94 RC21	46° 18.277'	129° 43.261'	7	5	50.0	14.6	1.24	10.6	0.21	8.0	12.6	2.44	0.10	0.11	99.8
95 RC27	46° 19.160'	129° 41.653'	7	5	49.8	14.6	1.59	11.1	0.18	7.5	11.7	2.78	0.13	0.14	99.5
95 RC28	46° 19.225'	129° 41.519'	7	5	49.8	14.7	1.58	11.1	0.20	7.4	11.7	2.83	0.14	0.14	99.6
95 RC25	46° 21.726'	129° 40.386'	7	5	50.3	14.1	1.74	11.9	0.22	7.0	11.6	2.84	0.16	0.14	99.8
95 RC24	46° 22.030'	129° 41.090'	7	5	50.6	14.3	1.65	11.2	0.22	7.1	11.7	2.79	0.20	0.17	99.8
95 RC23	46° 22.624'	129° 40.098'	7	5	50.4	14.4	1.34	10.6	0.22	7.9	12.1	2.55	0.09	0.13	99.8
94 RC1a	46° 22.664'	129° 40.009'	7	3	49.9	14.6	1.32	10.7	0.21	8.0	12.2	2.61	0.09	0.13	99.8
94 RC1b	46° 22.664'	129° 40.009'	7	2	50.1	14.4	1.41	10.9	0.20	7.7	12.3	2.59	0.10	0.12	99.8
95 RC22	46° 23.117'	129° 40.858'	7	5	49.5	15.4	1.19	10.6	0.20	8.4	12.1	2.33	0.11	0.08	99.9
94 RC2	46° 23.762'	129° 39.321'	7	5	50.0	14.5	1.37	10.8	0.21	7.8	12.3	2.60	0.09	0.13	99.8
94 RC3	46° 24.360'	129° 38.957'	7	5	50.1	14.3	1.43	10.9	0.22	7.7	12.3	2.63	0.09	0.13	99.8
94 RC4	46° 24.750'	129° 38.500'	7	5	50.4	14.4	1.50	11.0	0.20	7.4	11.9	2.81	0.11	0.14	99.7
95 RC21	46° 24.900'	129° 40.374'	7	5	50.1	14.3	1.68	11.8	0.22	7.1	11.4	2.90	0.17	0.18	99.8

*Standard data based on replicate analyses run over a several month period, each being the average of 5 or more spot analyses.

Table 2-2 continued.

Sample ID	Latitude N.	Longitude W.	Loc.	#	SiO ₂	Al ₂ O ₃	TiO ₂	FeO	MnO	MgO	CaO	Na ₂ O	K ₂ O	P ₂ O ₅	Total
94 RC5	46° 25.253'	129° 38.404'	7	5	50.1	14.3	1.40	11.0	0.19	7.7	12.4	2.62	0.09	0.14	99.9
94 RC6	46° 25.505'	129° 38.724'	7	5	50.2	14.5	1.41	10.6	0.20	7.9	12.3	2.64	0.09	0.11	99.9
94 RC23	46° 26.011'	129° 38.341'	7	5	49.6	14.8	1.42	10.6	0.20	8.0	12.3	2.64	0.11	0.13	99.7
94 RC7	46° 26.018'	129° 38.345'	7	5	50.0	14.6	1.39	10.6	0.20	7.9	12.3	2.69	0.09	0.10	99.9
94 RC8	46° 26.367'	129° 38.001'	7	5	50.5	14.2	1.51	11.4	0.21	7.3	11.7	2.78	0.12	0.12	99.8
94 RC9	46° 27.002'	129° 37.483'	7	5	49.8	14.7	1.58	11.2	0.21	7.7	11.7	2.76	0.11	0.13	99.8
94 RC10	46° 27.663'	129° 37.136'	7	5	50.2	14.4	1.50	10.9	0.22	7.6	12.1	2.72	0.09	0.12	99.8
94 RC11	46° 27.997'	129° 36.706'	7	5	49.6	15.4	1.10	9.7	0.18	8.5	12.7	2.59	0.05	0.08	99.8
94 RC12	46° 29.115'	129° 36.053'	7	5	50.2	15.0	1.10	9.7	0.17	8.2	12.8	2.57	0.07	0.07	99.8
94 RC14	46° 34.201'	129° 32.590'	7	5	49.6	15.6	1.20	9.9	0.19	8.4	12.3	2.58	0.07	0.07	99.9
95 RC32	46° 19.51	129° 46.255'	8	5	50.3	14.3	1.48	11.0	0.22	7.4	12.0	2.92	0.10	0.14	99.9
95 RC34	46° 22.418	129° 43.991'	8	5	49.9	14.9	1.39	10.0	0.21	7.9	12.2	2.83	0.07	0.11	99.6
95 RC35	46° 20.464	129° 45.991'	8	5	48.1	17.1	1.20	9.5	0.19	8.8	11.9	2.86	0.03	0.08	99.6
94 RC13	46° 34.359'	129° 35.227'	8	5	47.4	17.4	0.97	9.6	0.19	9.7	12.0	2.57	0.04	0.06	99.9
94 RC22	46° 18.474'	129° 46.687'	8	5	47.5	17.1	1.06	9.6	0.18	10.1	11.8	2.53	0.04	0.06	99.9
94 RC25	46° 02.231'	130° 00.712'	9*	5	49.9	14.3	1.54	10.9	0.22	7.2	12.2	2.90	0.20	0.15	99.4
94 RC26	46° 08.137'	129° 56.754'	9	5	49.5	14.4	1.55	11.3	0.21	7.3	12.2	2.95	0.19	0.17	99.7
94 RC28	46° 09.719'	129° 55.265'	9*	4	49.9	13.9	1.69	11.7	0.22	6.8	11.9	3.01	0.20	0.17	99.4
94 RC27	46° 09.646'	129° 56.253'	9	5	49.7	14.2	1.55	11.3	0.22	7.1	12.3	2.85	0.18	0.14	99.5
94 RC29	46° 10.384'	129° 55.394'	9	5	49.8	14.2	1.64	11.3	0.24	7.0	12.1	2.93	0.16	0.15	99.5
94 RC30	46° 10.260'	129° 54.716'	9	5	49.5	14.5	1.51	11.2	0.20	7.5	12.2	2.98	0.16	0.09	99.8
94 RC34a	46° 15.855'	129° 49.456'	9	3	49.5	14.4	1.54	11.4	0.17	7.3	12.1	2.99	0.17	0.13	99.8
99GTTA-1a	46° 09.720'	129° 55.230'	9*	3	49.8	14.1	1.67	11.8	0.20	6.9	11.9	3.05	0.21	0.15	99.7
94 RC32	46° 11.566'	129° 53.461'	10	5	48.2	17.1	1.15	9.0	0.18	9.2	11.8	2.81	0.02	0.07	99.6
94 RC33	46° 13.174'	129° 51.162'	10	5	48.1	17.3	1.12	8.7	0.15	9.3	12.3	2.78	0.07	0.09	99.9
62DR2-2	-46° 23'	~129° 28'	11	5	50.1	15.2	0.93	9.4	0.17	8.5	13.3	1.94	0.08	0.07	99.7
62DR2-4a	-46° 23'	~129° 28'	11	5	50.0	15.0	0.70	9.1	0.16	9.2	13.8	1.63	0.03	0.04	99.6
62DR2-6a	-46° 23'	~129° 28'	11	5	50.1	14.8	0.66	9.2	0.16	9.4	13.9	1.63	0.03	0.05	99.9
JdF-D2	average measured value		std	12 ^a	50.8	13.8	1.89	12.2	0.22	6.8	10.8	2.77	0.22	0.23	99.7
2392-9	average measured value		std	10 ^a	49.9	15.6	1.28	9.3	0.17	8.6	12.1	2.61	0.11	0.12	99.8
JdF-D2	2-Sigma Std. Dev. (relative %)				0.4	0.6	2.01	0.9	4.09	1.1	1.2	4.27	5.37	5.25	
2392-9	2-Sigma Std. Dev. (relative %)				0.6	1.1	3.17	0.9	7.92	1.1	1.0	1.71	5.61	11.03	
	Error used in plots (relative %)				0.5	1.5	3.0	1.5	10	1.5	1.5	5	5	10	

^aStandard data based on replicate analyses run over a several month period, each being the average of 5 or more spot analyses

USNM. Plots of SiO_2 abundance (wt.%) vs. Total (wt.%) of all oxides measured were made for each individual analytical run based on the secondary standard data acquired during that run. Where strong correlations existed, the raw abundance data for SiO_2 was corrected along a linear regression after Reynolds [1995]. Individual oxides were normalized to the standard JdF-D2 (unfractionated JdF-Reynolds, 1995) in order to greater facilitate their comparison to other published MORB data.

A small subset of the major element data was analyzed by EMP at the Smithsonian Institution by T. O'Hearn. Smithsonian data were not SiO_2 corrected but were normalized to the secondary standard VG2. In order to facilitate direct comparison to our microprobe data, these data were corrected to a "JdF-D2" normalization using numerous replicate samples and secondary standards analyzed in both labs. Measured values for the MORB standard JdF-D2 and long term analytical precisions are shown in Table 2-2. The complete major and trace element set is available from the author.

2.3.3 Trace Element Analyses

2.3.3.1 X-ray fluorescence

Seventy-five powdered basalt whole rock glasses were measured at the University of Florida for the elemental abundances of Ba, K, Nb, Rb, Sc, Sr, Ti, V, Y and Zr using a fully automated ARL8420+ x-ray fluorescence (XRF) spectrometer with a Rh-target end-window tube. Details of the operating conditions are reported in Smith et al., [1994]. Analytical precision based on long term reproducibility of standard reference materials is better than 5% for the elements K, Sr, Sc, Ti, V, Y and Zr. Precision is better than 10% for the elements Ba, Nb and Rb at values greater than 20, 3 and 2 parts per million (ppm) respectively. Precision drops to greater than 20% below these levels. Average values

and 2-sigma precisions for the standard BHVO-1 are given in Rubin et al., 1998, and average measured values for our internal standard ENDV (an enriched MORB from the Endeavour Segment, JdFR) are shown in Table 2-3.

2.3.3.2 ICP-AES/MS

Abundance data for major elements, the rare earth elements (REE) and selected other trace elements were measured by inductively coupled plasma source emission spectrometry, (ICP-ES) and inductively coupled plasma source mass spectrometry (ICP-MS) at the Geological Survey of Canada (GSC) in Ottawa. Elements/oxides measured by ICP-ES include major element oxides and a few trace elements (Co, Cr, Cu, Ni, and Zn), and those measured by ICP-MS include the REE and other trace elements. An additional group of samples recovered by wax core were analyzed by ICP-MS at the United States Geological Survey (USGS) in Denver. Fifty-two samples were analyzed by ICP for this study. A representative subset of these data are presented in Tables 2-3 (GSC data) and 2-4 (USGS data). Multiple internal standards analyzed by isotope dilution (ID) techniques on a thermal ionization mass spectrometer (TIMS) were included in each run to assess and correct for any inter-laboratory analytical bias. Correlation between the two laboratories was excellent. Details of analytical procedures, detection limits and precision for the GSC ICP data can be found on the internet at <http://132.156.95.172/chemistry/>, the GCS Analytical Laboratory homepage. Analytical procedures, detection limits and precision for USGS ICP-MS data can be found in Arbogast [1990].

Table 2-3 Concentration (ppm) of the Rare Earth and other Select Trace Elements in Representative

Submersible Samples Analyzed by ICP-AES ¹ , -MS ¹ , and XRF ² .										
Sample	*	2788-4R	2792-1R2	2794-2R	2983-R2	2983-R6	2792-4R	2792-5R	2792-7R	2948-11R
SiO ₂ ³	es	51.1	51.3	51.5	51.3	51.1	49.9	50.3	50.1	51.4
TiO ₂	es	1.3	1.26	1.3	1.31	1.33	1.51	1.57	1.55	1.22
Al ₂ O ₃	es	13.8	13.6	14	13.8	14.1	14.9	14.6	14.5	13.8
Fe ₂ O ₃ ⁴	es	12.8	12.5	12.9	12.9	13.2	12.5	12.7	12.5	11.6
Fe ₂ O ₃	es	1.1	1.2	1.8	0.9	1.5	1.4	1.1	0.7	1.1
FeO	es	10.6	10.2	10	10.8	10.5	10	10.5	10.6	9.4
MnO	es	0.19	0.19	0.19	0.19	0.2	0.18	0.19	0.18	0.18
MgO	es	7.14	7.05	7.18	7.2	7.39	7.4	7.35	7.23	7.37
CaO	es	10.9	10.7	11	11	11.2	10.5	10.4	10.3	11.5
Na ₂ O	es	2.44	2.52	2.63	2.56	2.54	2.67	2.66	2.57	2.43
K ₂ O	es	0.12	0.13	0.14	0.1	0.11	0.19	0.17	0.17	0.15
P ₂ O ₅	es	0.11	0.1	0.09	0.1	0.1	0.14	0.14	0.14	0.11
Total	es	98.8	98.2	99.8	99.3	100.0	98.8	98.9	98.1	98.7
Co	es	46	62	47	46	44	48	45	62	74
Cr	es	110	120	120	120	120	220	220	230	240
Cu	es	65	63	66	63	64	60	59	57	65
Ni	es	61	61	61	59	61	100	100	100	64
Zn	es	74	89	69	75	91	73	79	67	65
F	@	169	175	162	156	144	262	246	219	171
S	@	1263	1272	1257	1318	1305	1315	1256	1294	1151
V	xrf	405	377	396	406	401	378	386	388	332
Ba	xrf	32	25	25	29	28	36	33	36	29
Se	xrf	53	45	48	46	50	48	50	49	48
TiO ₂	xrf	1.54	1.49	1.58	1.54	1.53	1.82	1.89	1.88	1.39
K ₂ O	xrf	0.09	0.09	0.08	0.11	0.11	0.12	0.13	0.12	0.12
Cr	xrf	98	85	116	83	84	223	220	241	229
Zr	xrf	94	94	97	97	97	148	150	148	101
Sr	xrf	90	89	91	93	93	103	99	101	94
Y	xrf	38	38	39	40	40	51	50	51	35
Rb	xrf	1.1	1.7	1.4	2.1	1.4	1.6	1.6	1.7	2.0
Nb	xrf	3.5	3.4	3.2	4.1	3.0	4.5	5.1	4.7	4.7
Hf	ms	2.2	2.4	2.5	2.5	2.4	3.5	3.8	3.6	2.5
In	ms	0.11	0.12	0.13	0.09	0.13	0.1	0.14	0.11	0.09
Nb	ms	4	3.5	3.3	3.9	3.3	4.7	5.2	5.6	5.1
Rb	ms	1.3	1.3	1.5	1.2	1.3	1.7	1.9	1.8	1.8
Sn	ms	1.3	1.5	2.7	<0.5	0.7	1.7	1.9	1.1	0.9
Ta	ms	4.2	5.7	3.8	5	4.5	4.7	5.7	5	10
Th	ms	0.16	0.3	0.22	0.19	0.18	1	1.4	0.31	0.26
U	ms	0.08	0.07	0.1	0.1	0.07	0.14	0.13	0.12	0.11
Zr	ms	81	85	87	86	83	n.a.	n.a.	n.a.	92
La	ms	3.0	3.2	3.1	3.2	3.1	6.5	7.3	4.9	3.9
Ce	ms	9.1	9.4	9.3	9.4	9.2	17.0	18.0	15.0	11.0
Pr	ms	1.5	1.6	1.6	1.7	1.6	2.4	2.6	2.3	1.7
Nd	ms	9.2	9.2	9.4	9.7	9.6	14.0	15.0	14.0	10.0
Sm	ms	3.3	3.4	3.4	3.5	3.5	4.8	5.1	4.7	3.4
Eu	ms	1.3	1.2	1.3	1.3	1.3	1.5	1.6	1.6	1.3
Gd	ms	5.1	4.9	5.0	5.1	5.1	6.6	6.9	6.4	4.8
Tb	ms	0.90	0.94	0.97	0.95	0.94	1.20	1.20	1.20	0.84
Dv	ms	6.1	6.0	6.2	6.3	6.2	7.8	8.5	8.0	5.8
Ho	ms	1.3	1.3	1.3	1.4	1.4	1.8	1.9	1.8	1.3
Er	ms	4.0	4.0	4.1	4.0	4.0	5.0	5.2	5.2	3.7
Tm	ms	0.57	0.58	0.61	0.58	0.58	0.70	0.79	0.74	0.53
Yb	ms	3.8	3.6	3.9	3.9	3.8	4.9	5.2	5.0	3.7
Lu	ms	0.55	0.54	0.58	0.55	0.55	0.69	0.75	0.71	0.53
La/Sm(n)		0.57	0.59	0.57	0.58	0.56	0.85	0.90	0.66	0.72

¹ Analyzed at the GSC, Ottawa, ON.; ² analyzed at the Univ. of Florida, Gainesville, FL.; ³ values have been corrected for Si loss due to precipitation from analyzed solutions by extrapolation along a correlation between measured SiO₂ concentration and Total oxide abundance (r²=0.65) to a total of 99.7 wt%; *: method of analysis: @ by Dionex Ion Chromatography; es by ICP-AES, ms by ICP-MS, xrf by x-ray fluorescence; n.a. - not analyzed. All oxide concentrations are listed as wt % and all elemental concentrations are as ppm.

Table 2-3 continued.

Sample	*	2948-8R	2948-9R	2995-R2	2793-5R	2946-1R	2946-2R	2947-1R2	2948-10R	2948-1R
SiO ₂ ³	es	51.3	51.2	51.6	51.1	50.8	50.8	50.9	51.2	50.9
TiO ₂	es	1.23	1.19	1.25	1.15	1.19	1.21	1.19	1.1	1.03
Al ₂ O ₃	es	14	13.9	14	13.8	14.4	14.3	14.3	14.1	14.6
Fe ₂ O ₃ ¹	es	11.6	11.3	11.7	11.3	11.2	11	11.1	10.8	10.8
Fe ₂ O ₃	es	0.9	0.9	1	0.9	1.2	1.1	0.9	0.7	0.6
FeO	es	9.7	9.4	9.6	9.4	9	8.9	9.2	9.1	9.2
MnO	es	0.18	0.17	0.18	0.17	0.17	0.17	0.17	0.17	0.16
MgO	es	7.46	7.34	7.35	7.51	7.85	7.55	7.86	7.76	8.21
CaO	es	11.6	11.4	11.5	11.6	11.9	11.6	11.7	11.7	11.9
Na ₂ O	es	2.44	2.46	2.56	2.56	2.49	2.5	2.51	2.3	2.16
K ₂ O	es	0.14	0.14	0.16	0.1	0.11	0.1	0.11	0.09	0.1
P ₂ O ₅	es	0.11	0.11	0.12	0.09	0.08	0.09	0.08	0.09	0.09
Total	es	99.1	98.2	99.3	98.4	99.1	98.3	98.9	98.3	98.9
Co	es	41	42	40	47	46	52	42	40	48
Cr	es	240	250	230	200	300	290	310	270	320
Cu	es	66	67	66	70	67	63	68	66	72
Ni	es	66	66	64	68	85	80	90	78	94
Zn	es	66	65	64	64	67	69	67	63	60
F	@	160	152	167	144	150	145	160	136	125
S	@	1176	1190	1197	1154	1142	1102	1107	1130	1119
V	xrf	342	335	340	345	320	327	325	325	303
Ba	xrf	32	25	33	21	19	19	18	18	17
Sc	xrf	48	47	48	45	50	54	48	47	42
TiO ₂	xrf	1.41	1.39	1.43	1.33	1.34	1.42	1.36	1.27	1.16
K ₂ O	xrf	0.12	0.12	0.13	0.07	0.08	0.09	0.08	0.08	0.08
Cr	xrf	232	240	221	196	310	305	318	286	331
Zr	xrf	102	100	105	84	90	95	91	85	74
Sr	xrf	95	94	95	91	98	101	97	94	82
Y	xrf	36	36	37	33	34	34	33	32	29
Rb	xrf	2.1	2.1	1.7	1.0	1.2	1.4	0.8	1.3	1.5
Nb	xrf	4.2	4.6	4.5	2.6	2.9	3.6	3.1	3.2	4.0
Hf	ms	2.5	2.3	2.6	2.2	2.3	2.4	2.3	2	1.8
In	ms	0.09	0.11	0.11	0.11	0.09	0.11	0.09	0.08	0.1
Nb	ms	4.6	4.3	4.6	2.8	3.1	3.6	3.5	2.7	4
Rb	ms	1.7	1.8	1.9	0.82	0.83	0.88	0.87	0.85	1
Sn	ms	0.9	0.8	1	0.8	1.3	1	0.8	0.7	1.5
Ta	ms	4.9	4.7	5	5	5.9	9.7	8	3.3	11
Th	ms	0.27	0.24	0.25	0.2	0.13	0.14	0.18	0.16	0.15
U	ms	0.12	0.11	0.15	0.07	0.07	0.07	0.08	0.07	0.06
Zr	ms	90	89	99	79	87	88	87	73	69
La	ms	3.8	3.8	4.0	2.6	2.8	2.7	2.9	2.7	2.6
Ce	ms	11.0	11.0	11.0	7.9	8.5	8.4	8.9	8.2	7.5
Pr	ms	1.7	1.7	1.8	1.3	1.4	1.4	1.6	1.3	1.2
Nd	ms	10.0	9.7	10.0	8.2	9.0	9.0	9.2	8.4	7.7
Sm	ms	3.4	3.2	3.4	2.9	3.0	3.2	3.1	2.9	2.6
Eu	ms	1.3	1.2	1.3	1.2	1.1	1.2	1.2	1.1	0.9
Gd	ms	4.9	4.7	4.9	4.4	4.6	4.5	4.5	4.2	3.9
Tb	ms	0.90	0.87	0.92	0.80	0.80	0.87	0.82	0.76	0.69
Dv	ms	5.9	6.0	5.9	5.4	5.3	5.4	5.5	5.4	4.7
Ho	ms	1.3	1.3	1.3	1.2	1.2	1.2	1.2	1.1	1.1
Er	ms	3.7	3.6	3.9	3.4	3.4	3.5	3.5	3.3	3.1
Tm	ms	0.51	0.52	0.57	0.49	0.50	0.50	0.50	0.48	0.45
Yb	ms	3.6	3.5	3.6	3.3	3.3	3.5	3.4	3.3	3.0
Lu	ms	0.53	0.50	0.52	0.45	0.49	0.50	0.48	0.43	0.42
La/Sm(n)		0.70	0.75	0.74	0.56	0.59	0.53	0.59	0.59	0.63

¹ Analyzed at the GSC, Ottawa, ON.; ² analyzed at the Univ. of Florida, Gainesville, FL.; ³ values have been corrected for Si loss due to precipitation from analyzed solutions by extrapolation along a correlation between measured SiO₂ concentration and Total oxide abundance (r²=0.65) to a total of 99.7 wt%; *: method of analysis: @ by Dionex Ion Chromatography; es by ICP-AES, ms by ICP-MS, xrf by x-ray fluorescence; n.a. - not analyzed. All oxide concentrations are listed as wt.% and all elemental concentrations are as ppm.

Table 2-3 continued.

Sample	*	2948-2R	2948-6R	2948-7R	2949-1R	2949-4R	2788-6R	2792-1R1	2792-2R	2792-8R
SiO ₂ ³	es	n.a.	50.5	51.1	51.2	51.3	50.2	49.9	50.0	50.8
TiO ₂	es	n.a.	1.06	1.12	1.11	1.09	0.97	0.97	0.98	1.32
Al ₂ O ₃	es	n.a.	14.8	14.2	14.2	14.2	15.3	15.6	15.6	13.8
Fe ₂ O ₃ t	es	n.a.	11.1	11.2	11.4	11.2	10.2	10	10.3	11.6
Fe ₂ O ₃	es	n.a.	0.9	1.4	0.7	0.6	1.1	1	0.5	2.3
FeO	es	n.a.	9.2	8.9	9.6	9.5	8.2	8.1	8.8	8.3
MnO	es	n.a.	0.17	0.17	0.17	0.17	0.16	0.15	0.16	0.18
MgO	es	n.a.	8.33	7.62	7.4	7.29	8.32	8.48	8.36	7.43
CaO	es	n.a.	12.1	11.9	11.7	11.6	11.7	11.5	11.8	11.5
Na ₂ O	es	n.a.	2.21	2.31	2.41	2.36	2.43	2.59	2.59	2.53
K ₂ O	es	n.a.	0.12	0.11	0.08	0.1	0.06	0.07	0.07	0.11
P ₂ O ₅	es	n.a.	0.09	0.09	0.08	0.07	0.05	0.04	0.06	0.1
Total	es	n.a.	99.5	98.9	98.6	98.3	98.5	98.4	98.9	98.4
Co	es	170	46	52	40	70	45	47	49	49
Cr	es	250	310	240	220	220	310	300	310	280
Cu	es	67	72	64	67	67	82	77	80	61
Ni	es	61	94	66	60	62	110	120	110	72
Zn	es	61	62	65	66	64	53	47	53	61
F	@	n.a.	120	133	126	121	132	130	137	163
S	@	n.a.	1121	1126	1233	1200	1074	1020	1017	1179
V	xrf	330	308	331	340	328	263	248	268	353
Ba	xrf	19	21	16	17	21	14	10	13	22
Sc	xrf	51	45	48	51	47	51	42	49	53
TiO ₂	xrf	1.25	1.18	1.28	1.27	1.25	1.10	1.11	1.11	1.55
K ₂ O	xrf	0.07	0.08	0.08	0.07	0.07	0.05	0.04	0.05	0.08
Cr	xrf	227	328	245	225	224	354	319	345	301
Zr	xrf	78	75	83	80	80	77	78	79	102
Sr	xrf	83	81	91	85	86	95	97	95	103
Y	xrf	32	29	33	33	33	31	30	31	38
Rb	xrf	1.5	1.0	1.4	0.7	0.7	0.6	0.6	1.1	0.9
Nb	xrf	3.9	3.8	3.5	3.0	3.3	2.3	2.1	2.3	3.2
Hf	ms	1.9	2	1.9	2	2	1.8	2	2.1	2.4
In	ms	0.08	0.16	0.09	0.09	0.09	0.1	0.08	0.06	0.11
Nb	ms	4.2	3.6	3.3	2.6	2.9	1.7	1.7	2.2	3.6
Rb	ms	0.74	1	0.78	0.82	0.77	0.61	0.56	0.68	0.86
Sn	ms	1	3.3	0.5	0.8	0.7	2	1	1.7	1.1
Ta	ms	17	5.8	8.6	5.3	7.2	4.3	4.7	7.2	7.5
Th	ms	0.12	0.16	0.14	0.11	0.14	0.13	0.14	0.15	0.19
U	ms	0.06	0.07	0.07	0.09	0.07	0.05	0.07	0.06	0.07
Zr	ms	71	71	71	73	71	66	75	76	94
La	ms	2.4	2.7	2.5	2.4	2.4	2.1	2.0	2.3	3.1
Ce	ms	7.3	7.8	7.5	7.5	7.3	7.1	6.7	7.4	9.8
Pr	ms	1.3	1.3	1.2	1.3	1.2	1.2	1.1	1.3	1.7
Nd	ms	7.7	7.6	8.0	8.3	7.9	7.7	7.4	7.8	9.9
Sm	ms	2.8	2.7	2.8	2.9	2.7	2.7	2.6	2.8	3.5
Eu	ms	1.1	1.0	1.1	1.1	1.0	1.0	1.0	1.1	1.3
Gd	ms	4.2	3.9	4.2	4.3	4.1	4.0	4.1	4.3	5.0
Tb	ms	0.77	0.73	0.75	0.79	0.78	0.73	0.73	0.77	0.93
Dv	ms	5.1	4.8	5.1	5.3	5.5	4.8	4.6	4.8	6.3
Ho	ms	1.1	1.1	1.1	1.2	1.2	1.1	1.1	1.1	1.4
Er	ms	3.5	3.2	3.4	3.4	3.4	3.2	3.1	3.2	3.9
Tm	ms	0.48	0.44	0.45	0.49	0.45	0.46	0.43	0.46	0.55
Yb	ms	3.2	3.1	3.4	3.2	3.2	3.1	2.9	3.1	3.7
Lu	ms	0.45	0.43	0.45	0.46	0.45	0.43	0.43	0.45	0.55
La/Sm(n)		0.54	0.63	0.56	0.52	0.56	0.49	0.48	0.52	0.56

¹ Analyzed at the GSC, Ottawa, ON.; ² analyzed at the Univ. of Florida, Gainesville, FL.; ³ values have been corrected for Si loss due to precipitation from analyzed solutions by extrapolation along a correlation between measured SiO₂ concentration and Total oxide abundance (r²=0.65) to a total of 99.7 wt%.; *method of analysis: @ by Dionex Ion Chromatography; es by ICP-AES, ms by ICP-MS, xrf by x-ray fluorescence; n.a. - not analyzed. All oxide concentrations are listed as wt.% and all elemental concentrations are as ppm.

Table 2-3 continued.

Sample	*	2794-1R	2983-R1	2787-1R	2790-1R	2945-5R	99GTVA-1a	62DR2-1	62DR2-4a	Endv(n=4)	2-sigma
SiO ₂ ³	es	51.1	50.1	51.3	51.4	51.5	49.6	50.4	50.2	50.2	0.8
TiO ₂	es	1.33	1.2	1.24	1.21	1.25	1.35	0.62	0.6	1.39	0.02
Al ₂ O ₃	es	13.4	15.1	13.9	13.8	13.8	14.3	14.6	14.6	15.1	0.2
Fe ₂ O _{3t}	es	12.5	11	11.7	11.5	11.7	12.3	9.9	9.9	10.3	0.2
Fe ₂ O ₃	es	0.8	0.8	1.4	1.1	0.9	1.4	1.1	0.7	2.3	0.1
FeO	es	10.6	9.2	9.3	9.4	9.7	9.8	8	8.3	7.2	0.1
MnO	es	0.19	0.16	0.18	0.18	0.22	0.19	0.16	0.15	0.17	0.01
MgO	es	7	8.68	7.24	7.33	7.21	7.55	9.25	9.34	7.44	0.17
CaO	es	10.9	10.5	11.3	11.4	11.3	12	13.3	13.5	11.5	0.1
Na ₂ O	es	2.71	2.67	2.42	2.44	2.54	2.73	1.54	1.54	2.99	0.19
K ₂ O	es	0.1	0.14	0.13	0.15	0.14	0.22	<0.05	<0.05	0.24	0.01
P ₂ O ₅	es	0.09	0.11	0.1	0.11	0.11	0.12	0.04	0.04	0.15	0.01
Total	es	98.2	98.7	98.5	98.5	98.7	99.2	99.0	99.0	n.a.	n.a.
Co	es	43	89	47	67	49	53	44	44	34	1
Cr	es	130	300	180	190	180	270	460	450	265	6
Cu	es	62	77	69	66	67	87	100	100	65	1
Ni	es	57	160	62	61	62	43	93	93	76	1
Zn	es	69	62	69	65	70	54	45	47	59	2
F	@	169	209	166	173	167	196	<50	<50	218	13
S	@	1313	1093	1139	1196	1287	n.a.	n.a.	n.a.	n.a.	n.a.
V	xrf	399	278	352	350	356	336	n.a.	228	13	290
Ba	xrf	27	27	22	24	224	45	n.a.	16	2	81
Sc	xrf	48	43	50	49	49	46	n.a.	46	3	42
TiO ₂	xrf	1.53	1.40	1.44	1.40	1.46	1.56	n.a.	0.66	0.03	1.57
K ₂ O	xrf	0.09	0.11	0.11	0.11	0.12	0.19	n.a.	0.02	0.03	0.24
Cr	xrf	91	266	175	184	168	212	n.a.	426	25	236
Zr	xrf	94	146	98	94	100	n.a.	n.a.	22	4	119
Sr	xrf	90	115	99	100	124	n.a.	n.a.	29	9	165
Y	xrf	38	47	35	34	36	n.a.	n.a.	18	1	34
Rb	xrf	1.4	1.4	1.3	1.3	1.3	n.a.	n.a.	0.0	0.9	4.3
Nb	xrf	3.5	4.8	3.6	5.0	4.7	n.a.	n.a.	0.7	1.1	7.0
Hf	ms	2.5	3.6	2.3	2.3	2.4	3	1.6	1.4	2.8	0.1
In	ms	0.16	0.16	0.1	0.1	0.11	<0.05	<0.05	<0.05	0.13	0.04
Nb	ms	3	5.1	4.2	4.9	4.6	5.4	0.72	0.61	7.2	1.1
Rb	ms	1	1.4	1.3	1.2	1.6	2.2	0.29	0.23	4.4	0.1
Sn	ms	1	4.1	0.7	1.1	1	0.7	4.8	<0.5	2.5	1.4
Ta	ms	3.9	9.5	6.7	10	8.1	3.8	1.4	0.9	0.6	0.1
Th	ms	0.17	0.27	0.22	0.22	0.26	0.31	0.06	0.05	0.72	0.46
U	ms	0.08	0.1	0.11	0.11	0.13	0.14	0.06	0.04	0.18	0.01
Zr	ms	91	n.a.	87	84	94	n.a.	56	47	n.a.	n.a.
La	ms	2.9	4.4	3.5	3.4	3.6	4.8	0.70	0.70	6.8	1.0
Ce	ms	9.4	13	10	9.9	10	14	2.3	2.4	17	2
Pr	ms	1.6	2.2	1.7	1.6	1.7	2.0	0.45	0.43	2.4	0.1
Nd	ms	9.8	13.0	9.4	8.9	9.9	11	3.1	3.2	13	0
Sm	ms	3.6	4.5	3.0	3.1	3.3	3.3	1.3	1.4	3.8	0.1
Eu	ms	1.3	1.4	1.2	1.2	1.2	1.3	0.59	0.59	1.4	0.0
Gd	ms	5.1	6.0	4.8	4.4	4.7	4.6	2.4	2.5	5.0	0.1
Tb	ms	0.93	1.1	0.84	0.80	0.86	0.84	0.47	0.49	0.87	0.03
Dv	ms	6.5	7.1	5.6	5.4	5.9	5.2	3.2	3.2	5.4	0.1
Ho	ms	1.4	1.7	1.3	1.2	1.2	1.1	0.73	0.72	1.2	0.0
Er	ms	4.0	4.6	3.4	3.4	3.7	3.2	2.2	2.2	3.4	0.1
Tm	ms	0.59	0.63	0.54	0.49	0.53	0.45	0.32	0.31	0.46	0.01
Yb	ms	4.0	4.4	3.4	3.2	3.6	3.0	2.1	2.2	3.1	0.1
Lu	ms	0.56	0.65	0.47	0.48	0.53	0.47	0.33	0.35	0.44	0.01
La/Sm(n)		0.51	0.62	0.73	0.69	0.69	n.a.	n.a.	n.a.	n.a.	n.a.

¹Analyzed at the GSC, Ottawa, ON.; ²analyzed at the Univ. of Florida, Gainesville, FL.; ³values have been corrected for Si loss due to precipitation from analyzed solutions by extrapolation along a correlation between measured SiO₂ concentration and Total oxide abundance (r²=0.65) to a total of 99.7 wt%; *method of analysis: @ by Dionex Ion Chromatography; es by ICP-AES, ms by ICP-MS, xrf by x-ray fluorescence; n.a. - not analyzed. All oxide concentrations are listed as wt.% and all elemental concentrations are as ppm.

Table 2-4 Concentration (ppm) of the Rare Earth and other Select Trace Elements in Representative Rock Core Samples Analyzed by ICP-MS³

	2392-9	2352-2	94RC16	94RC20	94RC22	94RC25	94RC28	94RC30	94RC32	94RC34	95RC27	95RC28	95RC35	95RC36	95RC37	Chond. ⁴
La	3.0	3.7	3.3	4.1	1.3	4.8	4.8	3.6	3.4	5.2	4.4	5.2	2.1	4.4	4.9	0.367
Ce	9.3	11.5	10.8	12.7	4.9	13.8	13.8	11.0	10.2	15.0	12.7	13.8	7.5	12.7	13.8	0.957
Pr	1.5	1.9	1.7	1.8	0.9	1.9	1.9	1.7	1.6	2.2	2.0	2.1	1.3	1.8	1.9	0.137
Nd	8.0	10.0	9.6	9.7	5.2	10.0	10.0	9.2	8.8	12.0	11.0	11.0	7.4	9.6	10.0	0.711
Sm	2.8	3.6	3.7	3.6	2.1	3.4	3.4	3.3	3.3	4.1	3.8	3.8	2.7	3.3	3.5	0.231
Eu	1.1	1.3	1.4	1.3	0.9	1.3	1.2	1.3	1.2	1.5	1.4	1.4	1.1	1.2	1.3	0.087
Gd	4.0	4.9	5.3	4.8	3.0	4.6	4.5	4.4	4.4	5.3	5.3	5.1	3.8	4.3	4.6	0.306
Tb	0.7	0.9	1.0	0.9	0.6	0.8	0.8	0.8	0.8	1.0	1.0	1.0	0.7	0.8	0.8	0.058
Dy	4.8	5.9	6.4	5.9	3.8	5.2	5.2	5.1	4.9	6.0	6.3	6.1	4.6	4.9	5.2	0.381
Ho	1.0	1.3	1.4	1.3	0.8	1.1	1.1	1.1	1.1	1.3	1.4	1.3	1.0	1.0	1.1	0.085
Er	3.0	3.7	4.1	3.7	2.4	3.3	3.1	3.1	3.1	3.9	4.0	3.9	2.9	3.1	3.2	0.249
Tm	0.4	0.5	0.6	0.5	0.4	0.5	0.4	0.4	0.4	0.5	0.6	0.5	0.4	0.4	0.5	0.036
Yb	2.8	3.5	4.1	3.6	2.5	3.1	3.0	3.1	2.9	3.6	3.9	3.8	2.7	2.9	3.1	0.248
Ba	10	9	11	19	1	25	26	17	17	26	17	18	4	25	28	
Co	41	42	51	43	52	48	46	50	47	53	47	45	51	44	45	
Hf	2	3	2	2	1	2	2	2	2	3	3	3	2	2	2	
Nb	2.2	4.0	4.2	5.7	0.6	6.6	7.4	4.4	4.9	7.6	6.4	5.4	1.3	6.6	7.6	
Pb	3.2	2.1	5.3	<0.1	3.0	5.5	2.9	5.2	29.0	2.6	3.7	1.8	9.3	1.5	2.8	
Rb	0.8	1.0	1.1	2.1	<0.1	2.2	2.2	1.6	1.5	2.3	1.6	1.7	0.3	2.1	2.2	
Sc	39	41	53	44	41	46	46	44	42	55	47	46	38	43	45	
Sr	120	130	110	100	100	160	160	160	140	180	130	120	160	150	160	
Ta	0.1	0	0.2	0.2	<0.1	0.3	0.3	0.2	0.2	0.3	0.3	0.2	<0.1	0.3	0.3	
Th	0.1	0.10	0.1	0.2	<0.05	0.3	0.3	0.1	0.2	0.3	0.2	0.3	<0.05	0.2	0.3	
U	0.1	0.1	0.07	0.10	<0.05	0.10	0.20	0.07	0.08	0.10	0.1	0.1	<0.05	0.10	0.10	
V	280	340	440	370	240	350	350	310	320	400	380	380	250	330	340	
Y	28	34	39	35	23	31	30	30	30	36	37	37	27	28	29	
Zr	70	100	88	100	50	86	100	80	73	110	110	110	70	78	98	

³Analyzed at the USGS in Denver CO.; ⁴Chondrite normalizing values after Taylor, 1982.

2.3.4. Isotopic Analyses

Isotopic ratios of Sr and Nd were measured, on a representative subset of 30 and 18 samples respectively, at the University of Florida using a VG Isomass 354 thermal ionization mass spectrometer (Table 2-5). Natural glasses to be analyzed were dissolved at 120°C in sealed Teflon vials using distilled HF and a few drops of distilled HNO₃. Sr and rare earth elements were separated and collected using cation exchange chromatography (Dowex 50X12 resin), and Nd was separated from other rare earth elements (REE) using HCL elution on quartz columns packed with Teflon beads coated with bis-ethylhexyl phosphoric acid [after Richard et al., 1976]. Full procedural blanks are better than 0.2ng for Sr and 0.03ng for Nd.

Sr samples were loaded with Ta oxide on W single filaments. Sr ratio data were acquired in triple collector mode at a beam intensity of 2×10^{-11} A of ⁸⁸Sr, with corrections for instrumental discrimination made assuming ⁸⁶Sr/⁸⁸Sr=0.1194. Long-term measurement of NBS 987 yields a ⁸⁷Sr/⁸⁶Sr value of $.710216 \pm 45 \times 10^{-6}$ (2-sigma). Nd samples were run as metals loaded on one side of a triple filament array using a Re center filament and Ta side filaments. Data were acquired at a beam intensity of 5×10^{-12} A of ¹⁴⁴Nd, with corrections for instrumental discrimination made assuming ¹⁴⁶Nd/¹⁴⁴Nd=0.7219. Long term measurement of the La Jolla Nd standard yields a ¹⁴³Nd/¹⁴⁴Nd value of $.511834 \pm 18 \times 10^{-6}$ (2-sigma).

2.4. Regional CoAxial Segment Geochemical Characteristics

2.4.1 Major Element Trends

MORB recovered from the CoAxial Segment chemical attributes characteristic of “normal” incompatible element depleted MORB [Bryan et al., 1976; Sun et al., 1979;

Table 2-5. Sr and Nd Isotopes for Representative Samples.

Sample ID	Lat. N.	Long. W.	$^{87}\text{Sr}/^{86}\text{Sr}$	$^{143}\text{Nd}/^{144}\text{Nd}$	Location
221-0445	46.512	129.586	0.702267	0.513214	1
2670-4	46.526	129.579	0.702262	0.513248	1
2672-2	46.523	129.581	0.702263	0.513218	1
2792-3	46.532	129.574	0.702254	n.a.	1
2672-5	n.d.	n.d.	0.702325	0.513178	2
2672-6	46.521	129.573	0.702328	n.a.	2
2792-4	46.530	129.572	0.702300	n.a.	2
2985-1	46.524	129.569	0.702312	0.513179	2
2985-3	46.524	129.571	0.702306	0.513187	2
2948-11	46.300	129.720	0.702337	0.513188	3
2948-8r	46.291	129.724	0.702329	n.a.	3
2948-9r2	46.296	129.721	0.702331	n.a.	3
94rc20	46.302	129.719	0.702349	0.513127	3
221-0551	46.513	129.587	0.702266	0.513145	5
2671-2	46.509	129.589	0.702238	0.513195	5
2989-5	46.530	129.567	0.702304	n.a.	5
235-0700	46.274	129.744	0.702302	0.513215	4
2679-1T	46.313	129.704	0.702235	0.513192	4
2948-10	46.298	129.720	0.702273	0.513217	4
2995-3	46.345	129.560	0.702322	0.513165	4
2674-1	46.159	129.802	0.702390	n.a.	6
2675-2	46.178	129.797	0.702369	0.513112	6
2678-3	n.d.	n.d.	0.702314	n.a.	6
2945-5	46.156	129.810	0.702321	n.a.	6
94rc25	46.038	130.013	0.702501	0.513120	9*
94rc28	46.162	129.921	0.702499	0.513090	9*
99GTVA-1a	46.162	129.921	0.702505	n.a.	9*
62DR-1	46.387	129.473	0.702333	n.a.	11
62DR-2	46.387	129.473	0.702321	0.513199	11
62DR-4a	46.387	129.473	0.702308	n.a.	11

n.d. - no data; n.a. - not analyzed.

Schilling et al., 1983]. Figure 2-3 shows major element concentrations of 139 natural basaltic glasses from the CoAxial Segment analyzed by EMP. The basalt samples range in MgO content from approximately 9.6 to 6.5 weight percent (wt.%). Most of the chemical variability can be ascribed to the effects of fractional crystallization, but dispersion of data at a given MgO value is in excess of analytical uncertainty. Consequently, some variability in parental composition and/or conditions of crystallization is required.

Major element data from CoAxial lavas (Figure 2-3) form a field with a trend generally parallel to modeled low pressure fractional crystallization trends, with the exception of K₂O and P₂O₅ which increase in concentration at a greater rate than predicted by simple fractional crystallization models [Weaver and Langmuir, 1990]. The apparent greater enrichment of K₂O and P₂O₅ could be indicative of more complex crystallization processes such as *in situ* crystallization [Langmuir, 1989]. However, deviation from the modeled fractional crystallization trend also may be an artifact of the greater incompatible element depletions (relative to most CoAxial samples) present in the mafic sample used for the parent composition in the liquid line of descent (LLD) model shown in Figure 2-3.

Examination of geochemical data for MORB recovered from the CoAxial Segment and comparison to other data from other provinces of the southern and central Juan de Fuca Ridge reveals several notable relationships. Comparison of CoAxial lava chemistry to that of the Cleft Segment and Axial Seamount Province (which includes Axial Seamount caldera, north rift zone, and south rift zone) shows Coaxial lavas to be distinct in several ways. With the exception of one andesitic sample (not shown) recently

Figure 2-3: Variation of major element oxides (wt.%) with MgO in natural basaltic glasses from the CoAxial Segment, Juan de Fuca Ridge. Symbols are as shown in legend. Data for the 1998 Axial Seamount eruptive are after Perfit et al., 1998. Fields show compositional ranges for ridge axis basalts from the Cleft Segment (gray fill), Axial Seamount caldera (stippled fill), and Axial Seamount's north and south rift zones (white fill) for comparison. Fields are based on data from Smith et al., 1994 (Cleft Segment) and the unpublished data of Perfit et al. [in prep] (Axial Seamount). Lines show liquid lines of descent calculated for fractional crystallization at 1 kilobar (solid), 2.5 kilobars (dashed) and 5 kilobars (dotted) pressure using a recently modified version of the program of Weaver and Langmuir (1990). Parent composition used for the calculations was that of sample 221-0551, and older mafic basalt in contact with the 1993 eruption. See text for discussion.

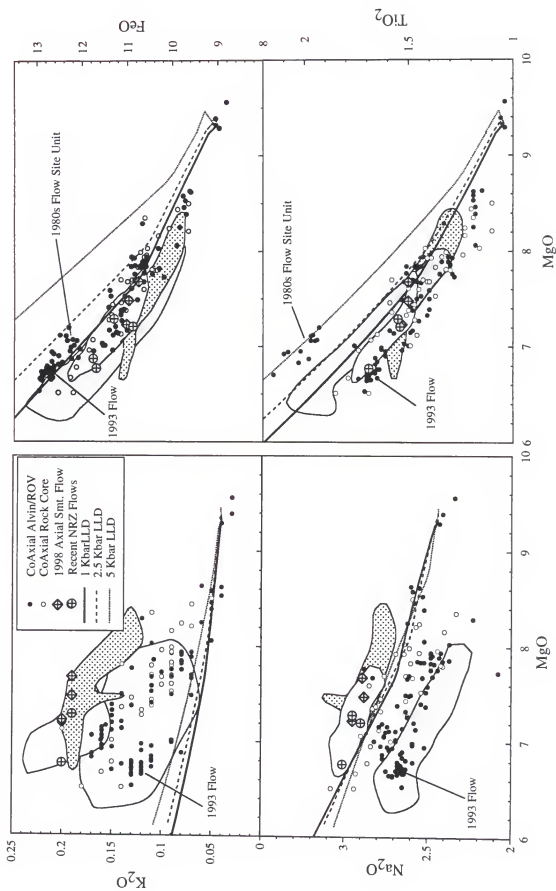
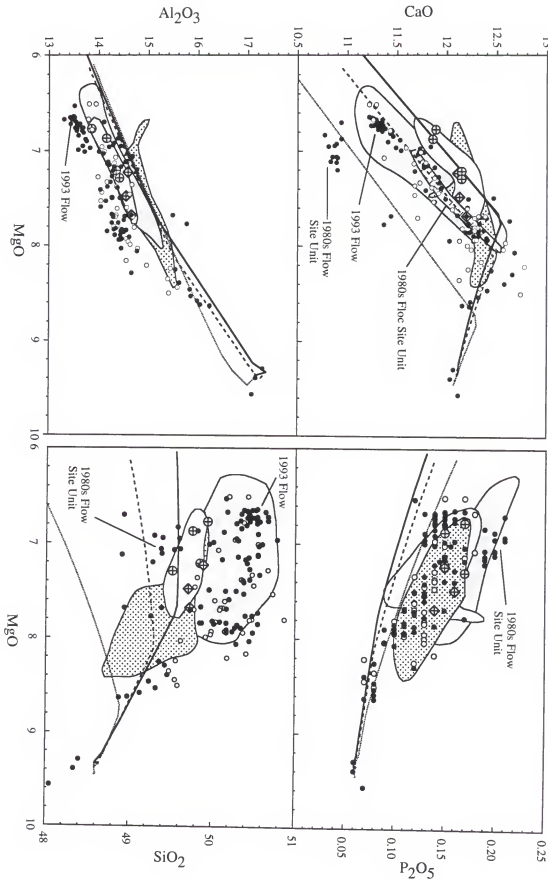


Fig. 3 continued



recovered from the North Rift Zone [Perfit et al., 1998] and some small on-axis volcanoes discussed below, glass data from the Axial Seamount display a more limited range in MgO, (~8.4 to 6.7wt%) than glasses from CoAxial (9.6 to 6.5wt%). Axial Seamount data from Rhodes et al. [1990] are not shown, but display a more restricted range in MgO content (6.9-7.7wt% MgO). In addition to having a greater range in MgO content, the CoAxial sample suite generally exhibits greater variation in other major element oxides at a given MgO than does the suite from Axial Seamount Province. It is, however, important to note that the Axial Seamount data set is smaller (81 glasses) than the CoAxial Segment database (139 glasses). As illustrated on MgO discrimination diagrams (Figure 2-3), there is significant overlap of CoAxial MORB with the field defined by Axial Seamount volcanics, but CoAxial lavas tend to have lower abundances of Al_2O_3 and CaO at a given value of MgO and extend to higher values of SiO_2 and FeO. Notably, less than ~10% of the CoAxial MORB have K_2O or Na_2O concentrations that fall within the Axial Seamount field.

Comparison of CoAxial MORB with MORB recovered from within the axis of the Cleft Segment, the southernmost segment on the Juan de Fuca Ridge [Smith et al., 1994] is useful for two reasons. Firstly, the Cleft segment represents one of the least morphologically and tectonically complex ridge segments on the Juan de Fuca spreading system, and it is somewhat spatially removed from complexities related to the Cobb melting anomaly. Secondly, the two MOR segments immediately adjacent to the CoAxial segment, the Vance segment to the south and the Cobb segment to the north, are very poorly petrologically characterized. Although data are very limited [Smith et al., 1994; Rhodes et al., 1990; Karsten et al., 1990; Delaney et al., 1981; our unpublished

data], the Cobb and Vance segments appear to have major element characteristics similar to that of the Cleft segment. Compared to Cleft samples, CoAxial MORB generally have lower abundances of Al_2O_3 and higher abundances of Na_2O at a comparable MgO abundance (Figure 2-3).

2.4.2 Isotopic and Trace Element Signals

Trace element concentrations in CoAxial Segment lavas have strong correlation to MgO suggesting that much of the trace element compositional variation observed within the CoAxial suite can be attributed to fractional crystallization and other differentiation processes. Additionally, covariation between trace elements that is inconsistent with crystallization or assimilation processes requires some variation in parental composition of lavas from this segment. In particular (as is discussed in greater detail in section 5.2.2), geochemical enrichments in a lava flow erupted at the Flow Site between 1981 and 1991 greatly extend the range of some trace element abundances, requiring differences in parental lava source characteristics.

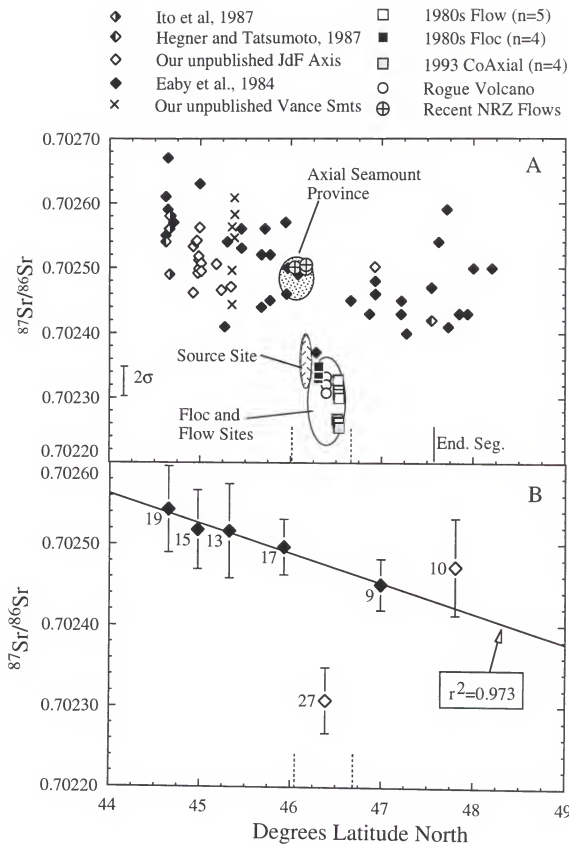
Discounting these few samples, the most incompatible enriched sample in the CoAxial Segment data field has an abundance of moderately incompatible elements (e.g. Y, Zr, Sm) 40% greater than the lowest abundance sample. Comparison of the most depleted and enriched CoAxial MORB shows that the abundances of the most highly incompatible elements (e.g. Nb, Ba, and Rb) are enriched by 250% to 300% relative to the lowest abundance sample (Table 2-3). Low-pressure fractional crystallization models [Weaver and Langmuir, 1990] indicate that ~25% fractional crystallization of the most MgO-rich sample is required to produce the range in MgO values observed in Table 2-3 (note that the range of MgO values in the sample subset shown in Table 2-3 is less than

that observed in the entire suite analyzed by EMP). Maximum corresponding enrichments in trace element concentrations (by assumption of a bulk distribution coefficient equal to 0) expected from fractional crystallization are limited to a factor of $\sim 1.3\times$ the parent composition when 75% of the original liquid remains. Consequently, over enrichment of the most incompatible elements relative to that expected from fractional crystallization are observed. Such incompatible element over enrichments in MORB suites (relative to fractional crystallization models) have been explained via more complex crystallization processes such as *in situ* crystallization [Langmuir, 1989], but may also be in part due to differences in primitive parental magma compositions.

Figure 2-4a shows latitudinal variation in the $^{87}\text{Sr}/^{86}\text{Sr}$ isotopic ratio of Juan de Fuca Ridge lavas. Data shown are a compilation of our data (Table 2-5 and unpublished data for Axial Seamount Province and the near ridge Vance Seamounts) and previously published data [Eaby et al., 1984; Ito et al., 1987; and Hegner and Tatsumoto, 1987]. $^{87}\text{Sr}/^{86}\text{Sr}$ data reported by Rhodes et al. [1990] (eight analyses) are not included in the compilation because they have values $10\text{-}20 \times 10^{-5}$ lower than the main data trend. This offset to of the Rhodes et al. [1990] data lower $^{87}\text{Sr}/^{86}\text{Sr}$ values is confirmed by Eaby et al. [1984] who report analyses of some of the same samples.

CoAxial MORB (latitudinal range indicated by vertical dashed lines in Figure 2-4) are clearly less radiogenic than MORB from the rest of the JdFR (and Pacific MORB in general). CoAxial Sr isotopic data ($n=25$) range from 0.70224 to 0.70239, whereas values for the rest of the Juan de Fuca are between 0.70240 and 0.70268. The less radiogenic nature of CoAxial $^{87}\text{Sr}/^{86}\text{Sr}$ relative to the rest of the JdFR is clearly illustrated in Figure 2-4b, which shows $^{87}\text{Sr}/^{86}\text{Sr}$ values averaged by region. The average CoAxial

Figure 2-4: (A) Distribution of $^{87}\text{Sr}/^{86}\text{Sr}$ for Juan de Fuca Ridge basalts plotted against latitude ($^{\circ}\text{N}$). Symbols and sources are as in the legend and text. Two-sigma error bar is shown at left. Vertical dashed lines at bottom mark CoAxial Segment boundaries, and the vertical solid line the Cobb offset. (B) $^{87}\text{Sr}/^{86}\text{Sr}$ compositions averaged by region. Error bars are ± 1 standard deviation for each average. Symbols from left to right are averages of data from: the southern Cleft Segment, the northern Cleft Segment, the Vance Segment (including Vance Seamounts), Axial Seamount Province (see text), the Coaxial Segment, the Cobb Segment and the Endeavour Segment. The number of analyses for each average are shown next to the data point. Line is a linear regression through the data points excluding the CoAxial and Endeavour Segments which respectively display anomalously depleted and enriched lavas chemistries relative to other JdF data. See text for discussion.



Segment $^{87}\text{Sr}/^{86}\text{Sr}$ value of 0.70231 is $> 15 \times 10^{-5}$ below the trend established by the other JdFR data south of the Cobb offset. It is interesting to note that although the overlap in data from the different regions is significant; there is a trend of decreasing $^{87}\text{Sr}/^{86}\text{Sr}$ with increasing latitude. This trend is also apparent in the unaveraged data (Figure 2-4a) suggesting that this trend is not simply an effect of the averaging of $^{87}\text{Sr}/^{86}\text{Sr}$. Regional surveys of JdFR Sr isotopes [Eaby et al., 1984, Rhodes et al., 1990] have noted that Axial Seamount does not have a Sr isotope signature distinct from the rest of the JdFR, but a regional trend in ridge axis lavas has not previously been recognized. It is important to note that while this trend is intriguing, it should be considered with caution. Significant overlap exists in the range of $^{87}\text{Sr}/^{86}\text{Sr}$ values observed in lavas from different regions of the JdFR, and latitudinal variation in $^{143}\text{Nd}/^{144}\text{Nd}$ does not display a similar linear correlation (though Nd data are more limited in quantity and spatial distribution).

CoAxial Segment lavas are not only distinct from other JdFR lavas with respect to Sr isotopic ratios, but also with respect to Sr abundance. Magmatic strontium abundances are not greatly affected by typical MORB crystallization (bulk liquid-crystal distribution coefficient is near unity), and in the absence of significant disequilibrium crystal accumulation or fractionation, large differences in Sr abundance between different lavas are likely a trait acquired during melt generation. A histogram of Sr abundances for MORB recovered from the Cleft Segment, CoAxial Segment, and Axial Seamount (Figure 2-5) shows the data-ranges for each of the three regions to be quite distinct from one another. As with the Sr isotopic ratios, CoAxial Segment lavas are the most depleted (in Sr abundance) on the JdFR. Only four samples from the CoAxial Segment (all among the oldest samples recovered) overlap with the data-range for Cleft Segment lavas, and

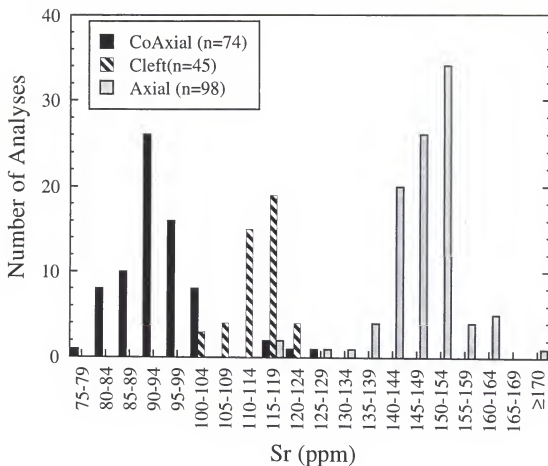


Figure 2-5: Histogram showing distribution of XRF analyzed strontium concentrations (ppm) in CoAxial Segment, Axial Seamount Province, and Cleft Segment basalts. Data used for the Cleft Segment is that of Smith et al., 1994. Data for Axial seamount province is a compilation of unpublished data of Perfit et. Al. (in prep) and that of Rhodes et al., 1990. See text for discussion.

only two Axial Seamount samples overlap with either the CoAxial or Cleft segment ranges. Strontium abundances of 96% of the CoAxial lavas analyzed fall between 79 and 104 ppm [average = 94 ppm], while 96% of Cleft Segment lavas have Sr between 105 and 121 ppm [average=114 ppm]. Axial seamount lavas have a Sr abundance range distinct from either two, with 97% of lavas having a Sr abundance between 130 and 170 [average=147 ppm]. Regional basalt isotopic and trace element signatures will be considered further below.

2.5. Discrimination of Discrete Flow Units

2.5.1 Visual and Bathymetric Flow Discrimination

Three different, recently erupted, mappable units have been identified on the CoAxial Segment. Assessment of differences in depth-data from repeat sea beam surveys of the segment clearly shows bathymetric anomalies corresponding to three different regions of very fresh lava mapped by submersible, ROV, deep camera tow, and AMS 60 deep tow sidescan sonar [Chadwick et al., 1995, Embley et al., in press]. Figure 2-2 shows AMS60 images for the Floc (2-2a) and Flow sites (2-2b) with outlines depicting the interpreted boundaries of the recent eruptives (after Embley et al. [in press], Figures 8 and 11). Superimposed on the sidescan images are the locations of basalt samples discussed below. Timing of the emplacement of these units is constrained to two intervals by the acquisition dates of the bathymetric data, the first occurring between 1981/82 and 1991 and the second occurring between 1991 and 1993. The lava associated with the SOSUS monitored T-phase seismicity [Embley et al., 1995] spatially corresponds with a sole bathymetric feature created between the 1991 and 1993 surveys, supporting this lava flow's association with the 1993 seismic event. This lava flow,

hereafter referred to as the 1993 flow, is approximately 3.8 km long, and up to ~500m wide and ~30m thick. Morphology of the 1993 flow is dominated by lobate to small pillowed forms. While most of the volume resides in the main body of the flow, several small segments of 1993 flow crop out north of the main body in a small graben that extends north and south of the eruption. This graben presumably served as the eruptive vent for the 1993 flow [Chadwick and Embley, 1998; Chadwick et al., 1995; Embley et al., in press]. Detailed descriptions of the individual flows and eruptive sites (Figures 2-1 and 2-2) are presented in Embley et al. [in press].

Comparisons between the 1981 and 1991 bathymetric data (Flow Site) and the 1982 and 1991 bathymetric data (Floc Site) reveal two additional positive bathymetric anomalies generated prior to the 1993 volcanic event. Seafloor observations confirm predictions of recent volcanic activity based on bathymetric differences between surveys, and identify two additional very fresh-looking volcanic constructional features. The first, hereafter referred to as 1980s Flow, is at the Flow Site approximately 700 meters east of the 1993 lava flow. It is of a similar morphology and size to the 1993 Flow [Chadwick et al., 1995]. The second, hereafter referred to as 1980s Floc, is at the Floc Site about 500 meters west of the zone of active venting and interpreted path of the 1993 dike emplacement [Embley et al., in press]. This flow (and associated bathymetric anomalies) consists of a linear array of several elongated mounds aligned with their long axis parallel to ridge strike. These fresh lavas occur over a distance of at least 7 km, and field relations indicate that these lavas were erupted along a pre-existing topographic ridge. While bathymetric anomalies were identified coincident with the thickest part of the recent flow, absence of bathymetric anomalies where glassy lavas were observed implies

that this flow is, in places, thinner than the 5-15 meter resolution of the bathymetric differencing technique [Embley et al., in press.]. The nearly complete overlap in the temporal ranges defined by the three surveys prior to 1993, 1981-1991 for the Flow Site and 1982-1991 for the Floc Site, limits the ability to discriminate whether these two older (relative to the 1993 flow), but still recent, lava flows were emplaced during one or two separate crustal accretion events. Although field observations of sediment cover suggest that the 1980s Floc eruption may be slightly older than the 1980s Flow eruption, these data also cannot preclude the possibility that both eruptive units are related to a single crustal accretion event. An examination of the inter- and intra- flow geochemical variation further constrains the genesis of these lavas and seafloor constructional features.

2.5.2 Geochemical Flow Discrimination

Examination of the data for the three recent CoAxial Segment lava flows, as shown in Figure 2-4a, reveals several important features. Firstly, for each of the eruptive units there is no statistically significant intra-flow variation in $^{87}\text{Sr}/^{86}\text{Sr}$ (or in $^{143}\text{Nd}/^{144}\text{Nd}$, not shown), although lavas from the segment as a whole do exhibit significant variation. Secondly, the 1980s Floc and Flow site eruptives cannot be distinguished from one another using $^{87}\text{Sr}/^{86}\text{Sr}$, although the 1993 Flow (number of samples analyzed (n) = 4) is distinctly less radiogenic than both the 1980s Floc ($n=4$) and 1980s Flow ($n=5$) eruptive units. There are no obvious crustal components or processes that could lower the Sr isotopic ratio of the 1980s magmas to produce the 1993 lava. Accordingly, the 1980s CoAxial lavas and the 1993 CoAxial lava cannot be easily related to a single parent, and are, therefore, neither comagmatic nor cogenetic. Further

discrimination of the three identified recent CoAxial eruptive units requires an examination of elemental abundances.

Electron microprobe data of natural glass samples recovered from each of the three recent eruptive units, along with average compositions and standard deviations for each unit, are shown in Table 2-1. Averaged data and analytical precisions for two MORB standards measured during the analytical runs are shown in Table 2-2.

2.5.2.1 The 1980s Floc Site Eruptive

Glasses from the three recent CoAxial flows exhibit both inter-flow and intra-flow major element chemical variability (Figure 2-6). The 1980s Floc Site eruptive, which is an ~7 km long linear array of elongate mounds, exhibits the greatest overall variation in major element concentrations. Data from samples collected *in situ* range from 7.5 to 7 wt.% MgO. The inclusion of wax core sample 94RC19, which targeted the southernmost mound of this flow (Figure 2-2a), extends the data to 8wt% MgO. This intra-flow variation, while not great (~12% relative to concentration), is well outside of analytical error, which is less than 2%. Perhaps more significant than the overall extent of intra-flow MgO variability is the co-variation between the major element oxides occurring in samples from this unit.

Figure 2-6a shows liquid lines of descent (LLDs) calculated for fractional crystallization at 1kbar pressure. A key to symbols is given in the figure caption. The models were generated using both the original program of Weaver and Langmuir (dashed line) [Weaver and Langmuir, 1990], and a more recently modified version of the program that includes the oxides K₂O and P₂O₅ in the calculations (solid line) [Reynolds, 1995]. Each of the algorithms of Weaver and Langmuir utilize a combination of experimental

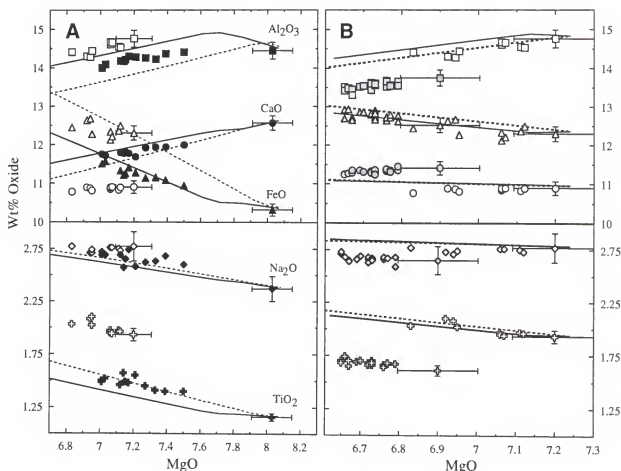


Figure 2-6: MgO concentrations plotted against concentration of Al₂O₃ (squares), FeO (triangles), CaO (circles), Na₂O (diamonds) and TiO₂ (crosses) for the 1980's Flow Site eruptive (open symbols), 1980's Floc Site eruptive (black filled symbols) and 1993 Flow Site eruptive (gray filled symbols). Error bars are shown for the highest MgO content sample in each of the three flows, and represent ± 2 standard deviations (2-sigma) of our standard data. In general, 2-sigma precisions (relative to concentration) are equal to or better than: 0.5% for SiO₂, 1.5% for FeO* (total Fe as FeO), MgO, CaO and Al₂O₃, 3% for TiO₂, 5% for Na₂O and K₂O, and 10% for P₂O₅ and MnO. (A) Comparison of glass data for the two flows erupted between 1981 and 1991. Trend lines show theoretical 1 kilobar liquid line of decent (LLD) calculations using the original program of Weaver and Langmuir, 1990 (dashed line) and a recently updated version of the same program designed to include K₂O and P₂O₅ (solid line) (See text). (B) Comparison of the two flows recently erupted at the Flow Site. Trend lines (generated using the updated LLD program) show the effects of pressure on modeled LLD paths with trend lines calculated at both 1 (solid) and 2.5 (dashed) kilobars pressure.

and thermodynamic data to predict the chemical evolutionary path, or liquid line of descent (LLD), of MORB liquids as they cool and crystallize mineral assemblages in chemical equilibrium with their host liquids. The algorithms are designed to model liquid chemical evolution during both closed system (equilibrium crystallization) and open system (fractional crystallization) crystallization processes. However, over the limited range of crystallization relevant to the intra-flow chemical variation observed, there are no significant differences between the LLDs generated using either the equilibrium or fractional crystallization models. The parent lava composition used for the calculations shown in Figure 2-6a is 94RC19, the southernmost and most mafic (MgO-rich) sample recovered from the 1980s Floc site eruptive. Five oxides plotted against MgO (Al_2O_3 , CaO, FeO^* , Na_2O and TiO_2) show covariation trajectories parallel to and bounded by the two model LLDs. The primary difference between the two models occurs because plagioclase's arrival onto the liquidus is suppressed in the model that considers effects of K_2O and P_2O_5 in calculating the LLD path.

The model shown in Figure 2-6a calculated using the original Weaver and Langmuir program (dashed line) has crystallization of clinopyroxene (cpx) as the sole liquidus phase beginning at a temperature of 1192°C . After only 4° of cooling (1188°C) both olivine (ol) and plagioclase (pl) begin crystallizing as the liquid becomes triply saturated. Using the same data to run the newer model (solid line) produces a liquid that once again begins with cpx as the sole liquidus phase, but crystallization now begins at a temperature 10 degrees higher (1202°C) than in the older (no K_2O or P_2O_5) model. Cpx remains the sole liquidus phase through 10 degrees of cooling at which point olivine joins the liquidus. Plagioclase begins crystallizing at the same temperature as in the older

model (1188°C), but by that time approximately 5% crystallization has already occurred. The effects of delayed plagioclase crystallization during the first 5% of crystallization are most dramatically seen in the LLDs for FeO (FeO enrichment is initially suppressed) and Al_2O_3 (Al_2O_3 shows greater initial enrichment). In general, the newer model does a better job of reproducing the natural variability observed in the 1980s Floc glasses, although our Al_2O_3 data suggest that initial increases in Al_2O_3 concentration are not as great as are predicted from the model. TiO_2 is the exception to the other major element oxides, showing better agreement with the older model.

The models' suggestion of clinopyroxene as a sole liquidus phase at low pressures in such relatively differentiated MORB is to be questioned. There is no experimental evidence that supports cpx occurring as a sole liquidus phase in such evolved compositions. Though small amounts of water may have the capability to suppress plagioclase crystallization [Michael and Chase, 1987], the algorithms of Weaver and Langmuir do not incorporate H_2O concentrations as a variable. Moreover, the 10 degree maximum interval of sole cpx crystallization is near the limit of experimental reproducibility [Grove et al., 1992]. It is, therefore, likely that sample 94RC19 was multiply saturated at its liquidus temperature.

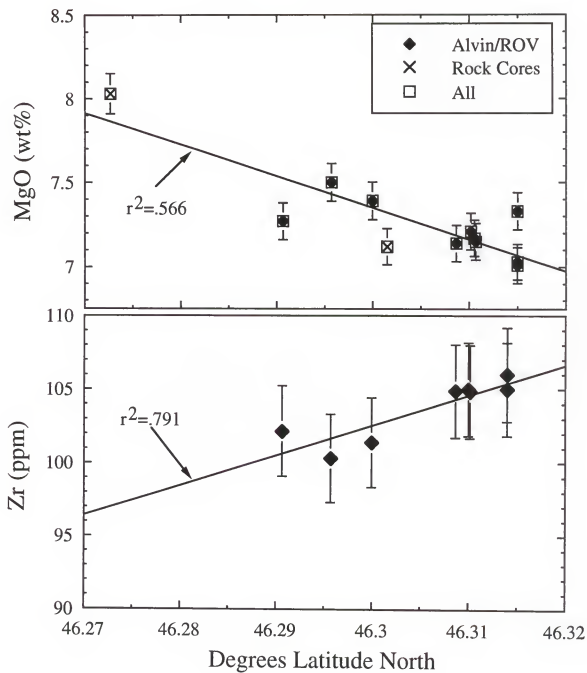
The close agreement of the data shown in Figure 2-6a with the predicted LLD trends suggests that elemental variations in glasses from the 1980s Floc eruption are consistent, within the limitations of current experimental models, with evolution along a liquid line of descent resulting from progressive cooling and crystallization. Thus, the chemistry of the sample recovered from the southernmost of the 1980s Floc mounds (94RC19) is consistent with it being cogenetic, comagmatic, and parental to the rest of

the samples from the 1980s Floc eruptive. Mineral phases observed in the lavas are consistent with the LLD models, and the total range of MgO compositions observed in the 1980's Floc Site eruptive corresponds to approximately 24° of cooling and 20% crystallization, as predicted by the LLD model (more recent version).

Trace element data are not from pure glasses, but instead are of "whole rock glasses" (which include a small amount of crystals which may not be a truly equilibrium assemblage). Accordingly, it is difficult to use our trace element data to rigorously constrain the crystallization conditions modeled, but it is nonetheless clear that some degree of liquid evolution has occurred within this flow unit. Whether this variation truly represents the process of fractional crystallization or is simply reflecting differences in the evolution of the liquid during equilibrium cooling and crystallization cannot be adequately assessed utilizing the glass data alone.

Trace element data gathered by XRF on whole rock glasses can be used to help constrain to what extent crystallization processes responsible for liquid differentiation within the 1980s Floc unit were closed (equilibrium crystallization) or open (fractional crystallization) system processes. If liquid differentiation observed within the glass data results from fractional crystallization processes, then composition of the whole rock glasses (crystals+liquid) should evolve with the liquid composition. Comparison of covariation of glass MgO and "whole rock glass" Zr concentration with latitude (Figure 2-7) reveals some interesting relationships. In this case, because the trend of the long axis of the 1980s Floc lava emplacement is aligned roughly parallel to ridge strike at ~022°, latitude is proportional to distance along strike of emplacement. A strong correlation exists between MgO content of glasses and distance along strike of the flow,

Figure 2-7: Latitudinal variation in MgO (upper) and Zr (lower) concentrations for basalt samples from the 1980s Floc Site eruptive. Symbols are keyed to method of recovery and are as shown in the legend. Error bars (± 2 -sigma) are shown, as are r^2 values for linear regression lines. Linear regressions are based on all the data shown. Though latitudinal increase in whole rock glass Zr concentrations (XRF) are similar to trends seen in the EMP data (upper) supporting the hypothesis of geochemical differentiation along the strike of flow emplacement, Zr increase is insufficient to be the result of pure fractional crystallization processes.



with MgO decreasing (cooling and crystallization increasing) from south to north. A similar, though less well defined, latitudinal covariation seems to exist in the Zr data, though the lack of “whole rock” data for the wax core 94RC19 limit the total variation observed. Nonetheless, if one extrapolates along the Zr trend (which over this limited range approximates a linear trend) to arrive at a Zr value for this sample, one would postulate it to have approximately 97 ppm Zr. Using this value as the parent composition, it becomes clear that the increase in Zr from 97 to 106 ppm (< 10% increase relative to concentration) is not sufficient to invoke pure fractional crystallization as the differentiation process. This small an increase would require a liquid-crystal bulk distribution coefficient (bulk D) of approximately 0.57, a value far too high for an element as incompatible during MORB crystallization as Zr (a “bulk D” equal to zero would yield a 25% increase in concentration after 20% crystallization.) Based on the variations of “whole rock” glass Zr data (and other highly incompatible trace elements not shown) it is likely that some intra-flow magmatic differentiation along the strike of emplacement exists. The differentiation process, however, was inefficient and cannot be accurately modeled via true Rayleigh fractionation [Shaw, 1970]. More likely, liquid differentiation was a process intermediate between closed system equilibrium crystallization and pure fractional crystallization.

2.5.2.2 The 1980s and 1993 Flow Site Eruptives

As discussed above, two discrete lava flows erupted between 1982 and 1993 have been identified at the Flow Site. Major element compositions of glasses recovered from the 1980s Flow unit (Figure 2-6 a and b, open symbols) and 1993 Flow unit (Figure 2-6b, gray filled symbols) show compositional variations similar to, but more restricted than,

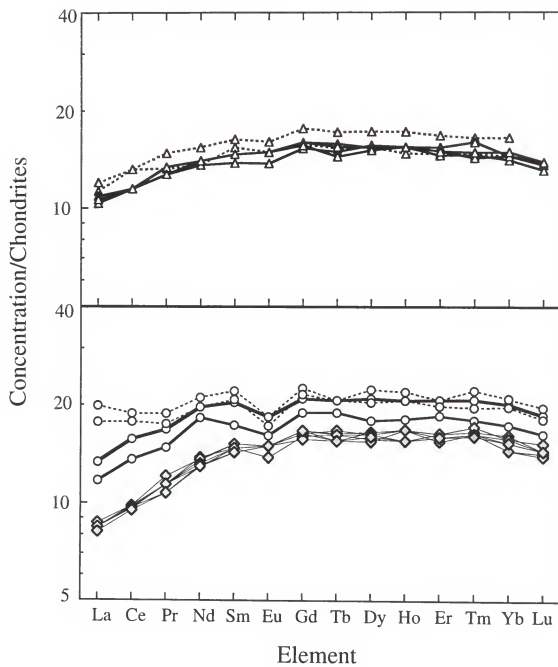
those from the 1980s Floc unit (shown only in Figure 2-6a). LLD models depicted by the trend lines in Figure 2-6b show the affects of varying crystallization pressure between 1 (solid line) and 2.5 kilobars (dashed line). In both models the starting composition is that of sample 2792-7, the most MgO-rich sample recovered from the 1980s Flow Site eruptive. Observed mineral phases and major element covariation with MgO in both of the Flow site units are consistent with LLD models and small amounts of liquid evolution ($\sim 9^\circ\text{C}$ of cooling and 2-5% crystallization). Although the total intra-flow chemical variation (4-5% relative to concentration for MgO) for these two units is close to analytical error, the consistent behavior between the different oxides and the LLD parallel trends are compelling. That being said, there is an apparent decrease in MgO and increase in Zr with latitude in the 1993 Flow (not shown), similar to the recent Floc Site eruptive, although only the northernmost and southernmost samples are completely distinguishable from one another considering the 2-sigma uncertainties. Neither of the two recent Flow Site eruptive units occurs over as great a distance along strike as does the 1980s Floc unit. The limited length of the Flow Site units, combined with lesser overall geochemical variability and restricted sample coverage for the 1980s Flow eruptive, inhibits the ability to define along strike chemical variability for these flows. Examination of Figure 2-6 does, however, clearly suggest that the two 1980s units are not cogenetic to one another, a conclusion that could not be drawn from the bathymetric or isotopic data.

Relative to the 1980s Floc unit, the 1980s Flow unit exhibits markedly higher TiO_2 and FeO^* and lower CaO . In fact, as can be seen in Figure 2-3, TiO_2 and CaO values for the 1980s Flow site unit appear to be rather anomalous compared to lavas from

the southern JdFR in general. Additionally, an examination of Table 2-3 reveals that, compared to the other recent CoAxial lavas that are both more and less geochemically differentiated, this basalt flow exhibits relatively high abundances in Zr, Hf, and to a lesser extent Nb. Elevated rare earth element (REE) abundances (relative to similarly differentiated samples from the segment) are also seen in samples from the 1980s Flow unit (Figure 2-8b, open circles). For example, REE patterns from the 1980s Flow unit are similar to, but have slightly higher chondrite-normalized abundances than those in either the 1993 Flow unit (Figure 2-8b, gray filled diamonds) or the 1980s Floc unit (Figure 2-8a). Abundances of the middle REE are about 20X chondrites, on average, in the 1980s Flow Site unit. That is approximately 30% higher than those from the Floc site unit (~15X chondrites) which has an average whole rock glass MgO content similar to that of the 1980s Flow unit. In addition to the elevated REE abundances, 1980s Flow samples have chondrite-normalized La/Sm (La/Sm_n) values (0.66-0.68) that are high relative to other Flow site lavas (0.48-0.62; $n=22$), but well within the CoAxial segment range in general (0.48-0.75, $n=52$).

Examination of the major and trace element data for the 1993 flow unit (Figures 2-6 and 2-8, Tables 2-1 and 2-3) reveal it to be the most depleted among the recent flows in incompatible elements, and among the most depleted along the entire CoAxial segment as well. While the 1993 flow is the most geochemically evolved (lowest MgO content) of the three recent CoAxial flows discussed, it has similar Na_2O contents to the more primitive 1980's lavas and the lowest overall concentrations of the elements Zr, Nb, Hf, Sr, Rb, K_2O , U and REE. Correspondingly, the 1993 flow samples are among the most light-REE depleted MORB recovered from the CoAxial Segment axis, having La/Sm_n

Figure 2-8: Chondrite-normalized (Taylor, 1982) rare earth element (REE) concentrations for MORB glasses from the 1980s Floc Site eruptive (upper figure; 6 samples analyzed), the 1980s Flow Site eruptive (lower fig.-open circles; 4 samples analyzed) and the 1993 Flow Site eruptive (lower fig.-gray triangles; 7 samples analyzed). All data were analyzed by ICP-MS at the Geological Survey of Canada with the exception of two wax core samples from the Floc Site (upper fig.-dashed lines) which was analyzed by ICM-MS at the USGS (Denver). Note the y-axis is an expanded exponential scale ranging from 5X to 40X chondrites. La and Ce are anomalously high in two of the 1980s Flow Site samples (lower figure-dashed lines), and may be suspect. Excellent intra-flow consistency is seen in the REE data, but variability occurs between different flows. Samples from the 1993 flow have the lowest overall REE concentrations and La/Sm_n (0.56-0.59), despite being the most geochemically evolved.



ratios between 0.56 and 0.59. In addition, as shown in Figure 2-4a, the 1993 eruptive also has the least radiogenic Sr isotopic signature of the three recent CoAxial lavas sampled. All of these observations suggest that the primary magma parental to the 1993 lava flow was more depleted in mantle magmatophile elements than other recent CoAxial lavas sampled.

2.6. Distinction of Regional Petrochemical Provinces

The comparison of minor and trace element characteristics of CoAxial lavas to those from other geologically identifiable terrains along the central JdFR suggests the CoAxial Segment is a magmatic province distinct from other segments of the JdFR. Further, examination of basalt geochemical data from the different geologically identified terrains allow us to test several hypotheses put forth by Embley et al. [in press] regarding their origin. In addition, these data can be used to investigate what relationship, if any, these central JdFR terrains have to the CoAxial Segment.

Firstly, it has been hypothesized that a highly reflective and unfractured area of seafloor south of $\sim 46^{\circ}18'N$ along Axial Seamount's NRZ delineates the extent of Axial Seamount related volcanism (boundary is shown as a solid black outline in Figure 2-1). Secondly, Embley et al. [in press] propose that the ridge north of $\sim 46^{\circ}18'N$, termed the western fault block ridge (WFBR), is not tectonically or magmatically related to Axial Seamount as has been previously suggested [Sohn et al., 1998]. Direct comparison of the geochemistry of historical CoAxial flows to the geochemistry of lavas recently erupted within Axial Seamount caldera, lavas recently erupted along Axial Seamount's north and south rift zones (including the 1998 eruption detailed in Embley et al., 1998; Perfit et al.,

1998 and others) [Smith et al., 1997], and lavas recovered along the WFBR allow for the evaluation of these two hypotheses.

2.6.1 Axial Seamount Province

Samples recovered from Axial Seamount NRZ and the WFBR are limited to those recovered by wax core and one video guided grab (Figure 2-1). NRZ samples (all within the region of highly reflective and unfractured sea floor) include 4 samples from a very fresh eruptive unit(s), as well as 9 other apparently older samples (Figure 2-1, Tables 2-2, 2-3, 2-4). With the exception of two samples recovered from small seamounts discussed in section 2.6.3, all of the samples recovered from this region bear strong geochemical affinity to samples recovered from within the caldera (Figures 2-3, 2-4, 2-9 and 2-10). Similarly, only one sample recovered from the SRZ [Perfit et al., 1998] shows any geochemical distinction from lavas recovered from within the caldera, and it too was recovered from a small seamount. Isotopically, all of the NRZ and SRZ samples analyzed to date fall within the field defined by samples from Axial Seamount caldera (Figure 2-4). The geochemical similarities between fresh lavas recovered from within Axial Seamount caldera and those recovered from the north and south rift zones extend beyond the Sr isotopic data.

Lavas sampled from Axial Seamount's rift zones are indistinguishable from caldera lavas with respect to the major elements (Figure 2-3), K_2O/TiO_2 ratios (Figure 2-9), REE (Figure 2-10) and other trace elements. The ratios of elements with similar distribution coefficients are unaffected by crystallization processes. Accordingly, regional differences in primitive magma composition are well illustrated by plots showing the ratio of two highly incompatible elements against latitude (e.g. Figure 2-9).

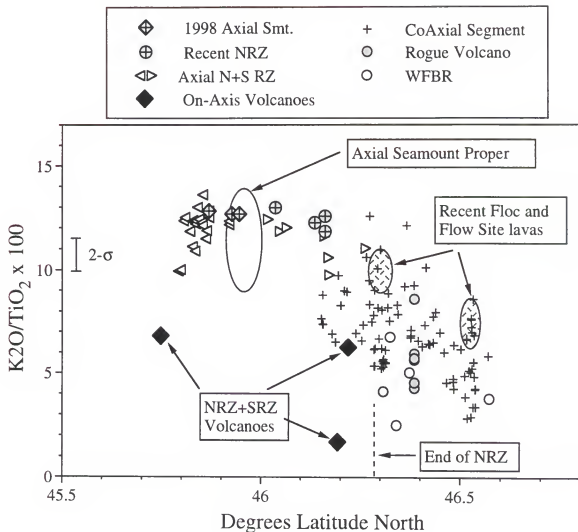
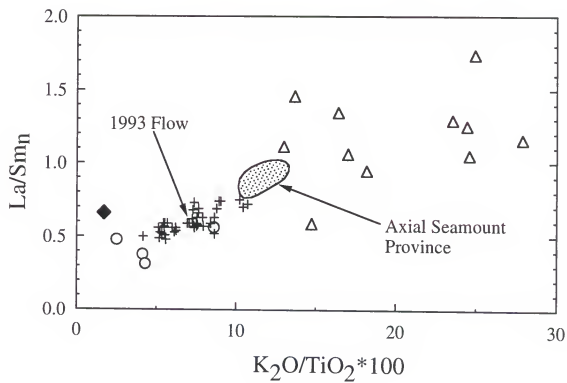
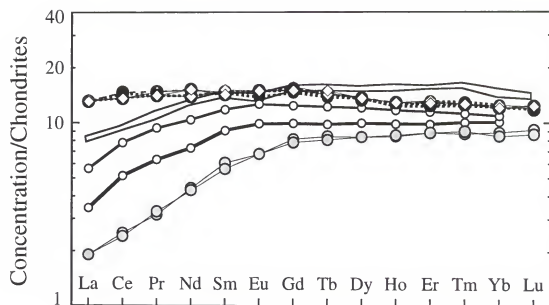


Figure 2-9: Regional variation in EMP K_2O/TiO_2 shown plotted against latitude. Symbols are as in the legend, and data fields are as labeled. All data are electron microprobe analyses of natural MORB glasses. Maximum possible variation in the ratio x/y due to analytical precision (2-sigma) governed by the equation $[(x+2\sigma_x)/(y-2\sigma_y)] - [(x-2\sigma_x)/(y+2\sigma_y)]$ equals 16% (shown for a $K_2O/TiO_2=10$). The field for Axial Seamount caldera, data points for its SRZ and the lava erupted in within the caldera and SRZ in early 1998 (Perfit et al., 1998) are all based on 73 unpublished EMP analyses (Perfit et al., in prep). The “Axial Proper” field encompasses data from all but one of the samples reported in Rhodes et al., 1990. Vertical dashed line marks NRZ boundary discussed in text.

Figure 2-10: (Upper) Chondrite-normalized (Taylor, 1982) REE patterns for regional basaltic lavas. Open diamonds show data for recent lavas from Axial Seamount caldera, and black-filled circles the recent NRZ flow discussed in text. Other symbols are as in Figure 2-9. Gray field marks the range of data shown in Figure 2-8 for the 1993 lava flow. Regional differences in LREE depletion are observed in lavas, with La/Sm ranging from ~0.9 in Axial Seamount province samples to ~0.3 in samples from Rogue Volcano. Samples from the WFBR (see text) have closer geochemical affinity to CoAxial Segment lavas than to lavas from Axial Seamount Province. (Lower) Covariation between EMP K_2O/TiO_2 and ICP La/Sm_n in regionally sampled lavas. Data from the Cobb-Eickelberg Seamount chain (Desonie and Duncan, 1990) are shown as open triangles. Strong positive correlation confirms that K_2O/TiO_2 in sampled lavas preserves a mantle derived geochemical signal and it is not dominated by crustal processes.



Due to the limited number wax core samples analyzed for trace elements by either XRF or ICP techniques, the ratio K_2O/TiO_2 (microprobe data) serves as the best available elemental parameter relatively unaffected by crystallization, to discriminate regional geochemical differences. Good regional correlation of K_2O/TiO_2 with La/Sm_n , Zr/Y and $^{87}Sr/^{86}Sr$ (Figures 2-10, 2-11 and 2-12) supports the assumption that the geochemical signal seen in K_2O/TiO_2 variations is related to differences in mantle-derived magmatic composition, and not to subsequent shallow level processes such as fractional crystallization and/or crustal assimilation. K_2O/TiO_2 values for Axial Seamount Province range from ~0.1 to 0.17, and samples recovered from the north and south rift zones (right and left pointing triangles respectively) are indistinguishable from samples recovered from Axial Seamount caldera.

Chondrite normalized REE patterns are similarly indistinguishable between the rift zone lavas and those from Axial Seamount caldera (Figure 2-10a). REE patterns from four samples of the youngest north and south rift zone lavas are shown (filled circles) and completely overlap in shape and abundance with data from the three caldera lavas analyzed for REE (open diamonds). As with the NRZ lavas for which data is shown, data for caldera lavas plotted are from the youngest identified (visually) in the caldera and have MgO contents comparable to the rift zone lavas. Seven fresh looking samples were analyzed from the Axial Seamount Province, and have a total range in La/Sm_n of 0.88-0.92. Consideration of the three older rift zone lavas (not shown) analyzed for REE extends the range of La/Sm_n down to 0.8. All of these observations, combined with the distinct chemistry of lavas recovered outside of the zone defined by Embley et al. [in press] (discussed below), support the hypothesis that the highly

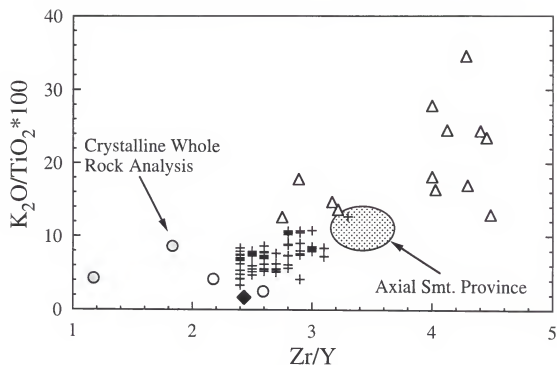
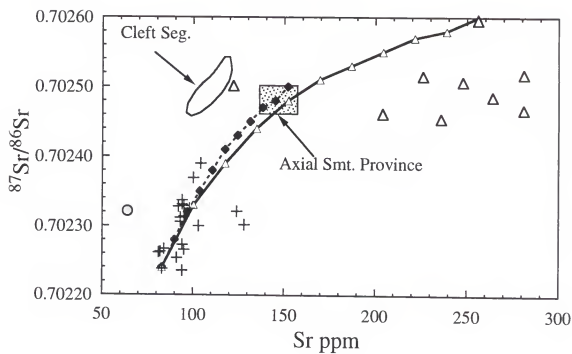
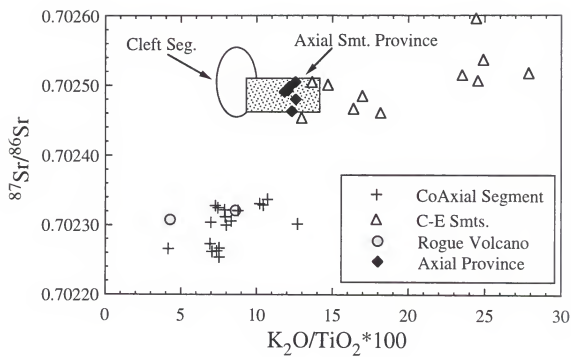


Figure 2-11: Regional covariation of MORB K_2O/TiO_2 (EMP) with Zr/Y (XRF/ICP). Strong positive correlation suggests that the assertions made in Figure 2-10 hold true for trace element groups other than the REE. Symbols and fields are as in Figures 2-9 and 2-10. See text for further discussion.

Figure 2-12: Correlation of the long lived radiogenic isotope $^{87}\text{Sr}/^{86}\text{Sr}$ with both $\text{K}_2\text{O}/\text{TiO}_2$ (upper) and Sr concentration (lower). Symbols are as in the legend. Fields for the Cleft Segment (grey field) and Axial Seamount Province (stippled field) are based on our unpublished data. Data for Axial Seamount samples analyzed for both isotopic and elemental abundances are limited, and, as such, the field shown encompasses the full range of values for all samples in our Axial Seamount database. The range is similar to other published data, with the exception of $^{87}\text{Sr}/^{86}\text{Sr}$ values reported in Rhodes et al. (1990) and discussed in text. Calculated mixing curves (Langmuir et al., 1978) are shown (lower) for mixes between a CoAxial Segment lava (221-0551) and lavas from Axial Seamount caldera (dashed line) and Warwick Seamount (DH4-4-solid line). Symbols on mixing-curves mark 10% mixing increments.



reflective and unfractured region of seafloor north of Axial Seamount caldera is blanketed with lavas that emanated from a magma source similar to the source that feeds Axial Seamount caldera.

2.6.2 The Western Fault Block Ridge

The topographic ridge extending north from NRZ (west of the CoAxial segment axial valley) becomes distinctly less acoustically reflective and more fissured north of approximately $46^{\circ} 18' \text{ N}$ latitude [Embley et al., in press]. This has been interpreted to reflect the greater age and sediment cover of the seafloor surface in this region, which apparently lacks recent volcanic overprinting of seafloor structures. Embley et al. [in press] conclude that the terrain between Axial caldera and $46^{\circ} 18' \text{ N}$ is related to recent constructional volcanism associated with NRZ volcanic activity, but north of $\sim 46^{\circ} 18'$ they propose the topographic ridge is unrelated to Axial Seamount magmatism and refer to it as the western fault block ridge (WFBR).

Five lava samples recovered by wax core from the WFBR are, on average, more MgO-rich than Axial Seamount Province lavas. These WFBR lavas have MgO concentrations from 10.1-7.4 (ave=8.8). These samples (shown as open circles on Figure 2-9) have $\text{K}_2\text{O}/\text{TiO}_2$ between 0.3 and 0.7 and fall outside of the field defined by Axial Seamount Province ($\text{K}_2\text{O}/\text{TiO}_2 = 0.1\text{-}0.17$). The low $\text{K}_2\text{O}/\text{TiO}_2$ values are more characteristic of the range defined by CoAxial Segment lavas (shown as crosses on Figure 2-9), for which 90% of 98 samples analyzed have a $\text{K}_2\text{O}/\text{TiO}_2 < 0.9$. Two of the WFBR lavas (94JdFRC22 and 95JdFRC35) have been analyzed for trace elements, and these data confirm their more incompatible element depleted nature relative to Axial Seamount Province lavas (Table 2-4). The two WFBR lavas have La/Sm_n values of 0.38

and 0.48 respectively compared to recent Axial Seamount lavas that range from 0.88-0.92. Zr/Y values of less than 2.6 in these samples are much more characteristic of CoAxial Segment lavas which average 2.7 and have 90% of samples less than 3.0 (101 out of 111 analyzed). By comparison, Axial Seamount lavas have Zr/Y greater than 3 (Figure 2-11).

While the sample base for the WFBR is small, and the two samples analyzed for trace elements do not uniformly exhibit the Sr elemental depletions characteristic of CoAxial lavas, these lavas nonetheless seem to bear a much stronger geochemical affinity to those from the CoAxial Segment than to those from Axial Seamount Province. WFBR lavas' elementally depleted nature relative to Axial Seamount lavas supports the hypotheses that the WFBR is a feature unrelated to Axial Seamount magmatism.

Samples recovered from the ridge that bounds the eastern side of the CoAxial Segment over a similar latitude range as the WFBR, are limited to six basalt samples from one dredge haul (95DR2). Though the data are limited, this topographic feature also seems to have geochemical signatures similar to both the CoAxial Segment and WFBR. These samples have XRF measured whole rock glass Zr/Y values between 2.4 and 2.8 (average = 2.6) compared with an average value of 2.7 for the CoAxial segment lavas. XRF K₂O/TiO₂ values are also similar to CoAxial, ranging between 0.054 and 0.096 (average = 0.077) compared with the average CoAxial value of 0.065.

2.6.3 Small NRZ Volcanoes

Two small seamounts within the NRZ terrain were sampled by rock (Figure 2-1, Tables 2-2 and 2-4, samples 94RC32 and 33) and both have very mafic compositions (greater than 9 wt.% MgO) and exhibit K₂O/TiO₂ values well below those measured in

lavas from Axial Seamount Province (Figure 2-9). A similar volcanic feature sampled from the Axial Seamount SRZ (Figure 2-9) also has similar mafic and depleted characteristics. Incompatible element depletions in these features appear not to be limited to the major elements. Sample 94RC32 was analyzed for trace elements (Table 2-4), and has both La/Sm_n and Zr/Y that are significantly lower than in lavas from Axial Seamount Province and comparable or lower than WRBF and CoAxial lavas (Figures 2-10 and 2-11). Two opposing hypotheses arise from consideration of the above observations and the data in Figures 2-9, 2-10 and 2-11. Firstly, these two small volcanoes may represent topographically high eruptive centers associated with the WFBR, and have survived subsequent burial by more recent volcanism along Axial Seamount NRZ. Alternatively, these two volcanic cones might have distinct magmatic plumbing systems similar to the many other small to medium sized near-ridge volcanoes that exist throughout the region (Figure 2-1, dark gray fill).

On the one hand, if the WFBR terrain is distinct from Axial Seamount Province as is suggested above and by Embley et al. [in press], then the presence of a volcanic feature geochemically similar to the NRZ volcanoes within the SRZ is difficult to explain via the first hypothesis. If, on the other hand, the alternate hypothesis is true and these small volcanoes have magmatic sources distinct from either Axial Seamount Province or the WFBR, then one might expect to see similar geochemical depletions in lavas from other near axis volcanoes in the region (Figure 2-1). Although there are numerous small near axis volcanoes in the CoAxial Segment region, particularly on the Pacific plate (Figure 2-1), only Rogue Volcano (on the JdF plate) has been sampled (Tables 2-2 and 2-3). Basalts analyzed from this volcano (6 for major elements, 2 for trace elements and Sr

isotopes) are all from one dredge (Figure 2-1). Similar to the small volcanoes sampled from the NRZ, lavas recovered from Rogue Volcano tend to be relatively MgO-rich and incompatible element depleted having average values for MgO (wt.%), Zr/Y, La/Sm_n and K₂O/TiO₂ equal to 9.1, 1.5, 0.33 and .058 respectively.

The association of near-axis volcanoes with unusually mafic and depleted basalt compositions has been previously recognized on the JdFR [Leybourne and Van Wagoner, 1991; Finney, 1989] and other regions of the MOR system [e.g., Fornari et al., 1988; Allan et al.]. In particular, lavas sampled from the Heck and Heckle Seamount chains (located west of the Endeavor Offset ~225 km north of the Rogue and NRZ volcanoes) have very similar geochemical characteristics to the lavas recovered from Rogue and the NRZ volcanoes. Similar to the samples discussed above, basaltic glasses recovered from the Heck and Heckle seamount chains [Leybourne and Van Wagoner, 1991] tend to be mafic, averaging greater than 8.4wt% MgO and reaching concentrations as high as 9.3 wt.%. Heck and Heckle lavas are also depleted in incompatible elements, having Zr/Y less than 2.4, K₂O/TiO₂ less than 6 and La/Sm_n less than 0.42. The above elemental considerations support a hypothesis whereby the small NRZ seamounts result from magmatic activity unrelated to either Axial Seamount or the WFBR, possibly having origins similar to other near-axis volcanoes in the central JdFR region.

2.6.4 The CoAxial Segment

CoAxial Segment lavas are perhaps most uniquely characterized by their depleted strontium abundances and isotopic ratios relative to other axial JdFR basalts (Figures 2-4 and 2-5). CoAxial segment lavas also tend to be quite incompatible element-depleted having Zr/Y, K₂O/TiO₂ and La/Sm ratios ranging from 2.4-3.2, 0.03-0.1, and 0.48-0.75

and averaging 2.7, 0.07 and 0.6 respectively. Lavas from the 1980s Floc unit and the Source site (though not well represented in our sample suite) tend to be the least LREE depleted of any CoAxial lavas analyzed, consistently having the highest La/Sm_n (greater than 0.68). These samples also tend to have among the most radiogenic Sr isotopic signatures observed for the segment. With the exception of the 1980s Flow unit discussed in detail in section 5.2.2, samples from the flow site tend to be the most radiogenically and elementally depleted along the segment. This effect is exemplified by a trend of generally decreasing $\text{K}_2\text{O/TiO}_2$ with increasing latitude along the segment (Figure 2-9). Although CoAxial lavas are elemental-depleted, their compositions do not display the extreme of elemental depletions evident in the “near-axis” volcanoes discussed above (Figures 2-9, 2-10 and 2-11).

2.7. Discussion

2.7.1 Recent CoAxial Segment Volcanism

The volcanic eruption that occurred on the CoAxial Segment in 1993 at the Flow Site was a phenomenon unique in the study of mid-ocean ridges because it was an *actively monitored* crustal accretion event [Fox et al., 1995]. Real time seismic monitoring of the JdFR recorded the dike event that fed the 1993 eruption. The locations and migration of the epicenters of T-phase seismicity recorded during the event led to the initial conclusion that the dike intruded along strike and likely originated from the north rift zone of Axial Volcano [Dziak et al., 1995]. Although it was recognized that topographic effects could have caused a westward bias in the locations of epicenters, Axial Seamount could not be ruled out as a source for the eruption. Subsequent studies of post-emplacement seismicity on the CoAxial Segment concluded that the dike path

was along the CoAxial spreading center axis, and suggested that T-phase epicenters associated with the 1993 seismic event were mislocated [Sohn et al., 1998]. The documented geochemical differences between Axial Seamount Province lavas and CoAxial Segment lavas in general, and the 1993 flow specifically, make it highly unlikely that Axial Seamount served as the source for the 1993 dike. The geochemical distinction of the two regions is consistently seen in the major element, trace element and isotopic characteristics of lavas from each region. A “CoAxial” origin for the 1993 eruption is further supported by geologic and hydrologic observations [Embley et al., in press], making a compelling argument against Axial Seamount as a source for the 1993 event.

Two other eruptive units have been emplaced along the CoAxial Segment since 1981 [Embley et al., in press]. Differences in the major and trace element geochemistry between these two flows strongly suggest that lavas from these two units are neither comagmatic nor are they cogenetic. This leads to the conclusion that at least three separate eruptive events have occurred along this ridge segment during a span of twelve years or less. This is the first such area of the MOR system where such detailed constraints on eruptive frequency and distribution have been established, allowing for a more detailed investigation of the temporal and spatial aspects of crustal accretion at this medium rate spreading center than would otherwise be possible. Relative location of the three lava flows establishes the “neovolcanic zone” for this ridge segment to be greater than 1 km wide at the decadal time scale and over lengths of less than 25 km along strike.

The eruption of the 1980s Floc unit along a preexisting constructional ridge [Embley et al., in press] and the *in situ* sampling of relatively fresh and geochemically

distinct lava from a kipuka in the 1980s Floc unit (sample 2948-10R) is intriguing. The presence of relatively fresh, non comagmatic lavas comprising the pre-existing ridge suggests that at least two eruptions have occurred at this location in the recent past (less than 100's of years), some 500m west of the interpreted ridge-axis and locus of current geologic and hydrothermal activity. Accordingly, although CoAxial eruptive activity might not be as well focussed as in some other regions of the Juan de Fuca Ridge and northern East Pacific Rise studied in comparable detail [Embley and Chadwick, 1994; Fornari et al., 1998], there is likely some control of preexisting structure on the location of dike emplacement along this segment. This is consistent with the observation that the 1993 dike was emplaced along the trend of a preexisting fissure system that was apparently reactivated during the intrusive event.

Perhaps as intriguing as the temporal/spatial variation in lava emplacement is the variation observed in lava chemistry. None of the flows which have been emplaced within the CoAxial Segment axis over a twelve year period appear to be cogenetic and each has a genetic origin uniquely distinguished from the other by elemental concentrations and ratios and/or Sr isotopic ratios. In addition, it has been noted that samples recovered from the CoAxial Segment axis show a greater dispersion in elemental concentrations at a given MgO value than do other Segments of the southern JdFR. These observations suggest that there is not a well-mixed, long-lived magma chamber serving as the source for CoAxial Segment volcanism. Inter-flow geochemical variability is more suggestive that the magmatic plumbing underneath the CoAxial Segment is poorly developed, with individual melt bodies being relatively small and short lived. In fact, differences in mantle derived geochemical signals observed between the two recent

units at the Flow site (Sr and Nd isotopes and the ratios of elements highly incompatible with mantle mineral assemblages) show that mantle geochemical heterogeneity can be preserved by seafloor crustal accretion here on temporal and spatial scales of a decade and 500m.

The hypothesis that CoAxial Segment volcanic activity (at least that part north of the poorly geochemically characterized Source site) is not related to a large, long-lived, controlling magma chamber is consistent with the observed geology and geophysics. Seismic data for the upper three km of the crust show no evidence of a low velocity zone associated with an axial magma chamber reflector [Sohn et al., 1997], and seafloor compliance measurements conducted on the segment failed to detect any magma bodies in the crust that comprises the CoAxial Segment axial valley [Crawford, 1994]. Additionally, Sohn et al. [1997] have interpreted the crustal seismic structure of the northern portion of the CoAxial Segment in terms of a dominantly amagmatic extensional terrain.

The CoAxial Segment's morphology and structure is not like that of some magmatically robust segments such as the Cleft segment to its south [Embley et al., 1991] or the N. East Pacific Rise (EPR) [Haymon et al., 1993]. Along much of the Cleft Segment and the 9°-10°N segment of the EPR, axial volcanism appears to be dominantly focussed within a relatively narrow region. These areas, presumably underlain by well developed magma chambers [Smith et al., 1994; Fornari et al., 1998], are associated with phenomenon such as high temperature hydrothermal venting, formation of lava lakes and large regions of volcanic collapse [Embley et al., 1991; Haymon et al., 1993; Fornari et al., 1998; Kent et al., 1993]. None of these features have been observed in the northern

portion (Floc and Flow Sites) of the CoAxial Segment. CoAxial Segment volcanism occurs over a zone greater than 1 km wide, and significant portions of the CoAxial terrain north of the Source Site are highly fissured giving the impression that amagmatic extension has played an important role in the development of axial structure, despite the evidence for recent volcanic activity.

The spatial patterns of recent volcanism and abundant fissures within the neovolcanic zone of the northern CoAxial segment are consistent with seismic data that suggest no shallow magma chamber underlies this portion of the segment. These observations are consistent with a southerly source for recent volcanism, and support the hypothesized lateral propagation of the 1993 dike, thought to have originated south of the Floc Site.

2.7.2 Intra-flow Geochemical Variability.

Although no direct evidence for crustal magma bodies exists, it is clear that the recently erupted CoAxial lava flows have undergone substantial differentiation prior to eruption. One situation where chemical fractionation could have taken place is during the lateral emplacement of the dike through the crust. In the case of the 1993 eruption, the dike is thought to have migrated some 40 km along axis [Dziak et al., 1995] over a period of ~2 days. While evidence of intra-flow fractional crystallization in the 1980s Floc unit along the strike of its emplacement (Figure 2-7) is intriguing, a closer examination of major element data shows this effect to be insufficient to generate the observed fractionated compositions of the 1993 lava flow. Basaltic glasses show a nearly 1 wt.% variation in MgO over the 7 km long axis of the flow. This change in MgO corresponds to ~20% fractional crystallization of the most mafic portion of that flow based on LLD

calculations. On the surface then, it would seem that a fractional crystallization rate of ~3% per km would be sufficient, over a lateral emplacement distance of ~40km, to account for the approximately 75% fractional crystallization required to differentiate a moderately primitive MORB magma (~9.5wt% MgO) to the composition of the 1993 Flow (~6.7wt% MgO). Two observations argue that chemical differentiation of the 1993 flow could not have entirely occurred during dike intrusion and eruption.

Firstly, comparison of the intra-flow whole rock glass data to the EMP data for the 1980s Floc unit shows that the liquid differentiation process was nowhere near as efficient (by ~50%) as pure fractional crystallization predicts. In addition, subaerial analogues to the 1993 diking event do not exhibit this degree of magmatic differentiation along the path of the dike injection. This is not surprising, as it is difficult to envision efficient separation of crystals from liquid in the dynamic environment of a dike emplacement that occurs over the course of only a few days. Nonetheless, some bulk chemical differentiation does appear to have occurred along the 7 km strike of the 1980s Floc Site flow.

Secondly, the petrography of samples from the 1993 flow suggests that the magma likely had some residence in a crustal magma body prior to dike intrusion. Glassy rims of the 1993 lava samples tend to be quite crystalline (visual estimates of up to 10-20 modal percent). Much of the crystal assemblage consists of small, quenched microphenocrysts of plagioclase and clinopyroxene consistent with a phase of rapid cooling and crystallization prior to eruption. There are, however, a second population of larger phenocrysts (up to 2 mm) of olivine, clinopyroxene and plagioclase, some of which show signs of resorption and chemical disequilibrium. It is unlikely that this

second population of larger grains could have formed during the time frame of the intrusion event, and it is more likely that they crystallized from a cooling crustal magma body. This conclusion in no way, however, detracts from the importance of the intra-flow chemical variation that does exist within the 1980s Floc unit, but does infer the presence of a CoAxial crustal magma body that has yet to be imaged.

Recently a positive correlation between intra-flow chemical variability and eruption volume has been suggested [Rubin et al., 1998; Perfit and Chadwick, 1998]. Additionally, others have suggested an inverse correlation between spreading rate and eruptive flow volume [Sinton, 1997; Perfit and Chadwick, 1998]. The strong correlation of glass geochemistry along the strike of emplacement for the 1980s Floc unit, and to a lesser extent the 1993 Flow unit, may imply a causal mechanism relating these two observations. Intra-flow geochemical variation observed in the three recent CoAxial lava flows are all consistent with small amounts of magmatic differentiation along low pressure LLDs. This is most easily explained via small degrees of cooling and crystallization occurring during lava emplacement. This hypothesis is supported by the petrography of the 1993 CoAxial flow which has a large proportion of microlites and microphenocrysts in the outer most glassy selvage of the lava flow, a characteristic that is likely related to a rapid phase of cooling prior to eruption. The elongate lineated nature of the 1980s Floc unit and 1993 Flow unit, their systematic intra-flow geochemical variation along ridge strike, and the observation of lateral dike propagation during the 1993 event all suggest that the 1980s Floc unit was emplaced via a similar south to north propagating dike. Does the process of lateral propagation of intruding (and eventually erupting) dikes in the shallow ocean crust aid in the generation of intra-flow geochemical

heterogeneity? These data would support that conclusion. Is a dike more likely to propagate long distances laterally in a slower spreading environment? Intuitively it would seem that an increasingly tensile stress regime would promote such propagation, but factors such as ridge-axis structure, magma density, eruptive pressure and topographic considerations are all factors to be considered. It is intriguing that several investigations of JdFR volcanism have led to conclude that lateral dike propagation served as the mechanism of lava emplacement [Embley et al., 1991 and in press]. Conversely, investigations of the northern East Pacific Rise have concluded that recent eruptions were emplaced through mostly vertical dike intrusion [Wright et al, 1995].

2.7.3 Regional Petrochemical Characteristics

This study recognizes four petrochemically distinct magmatic provinces in the Axial Seamount-CoAxial Segment region. Axial Seamount Province and its south and north rift zones comprise one. Another is comprised of the western fault block ridge. The third is the Coaxial Segment, and the last is comprised of several small axial and near axis volcanoes.

2.7.3.1 Axial Seamount

Samples recovered from Axial Seamount north and south rift zones are generally geochemically indistinguishable from lavas recovered from within the caldera. Data presented here support the boundaries of NRZ volcanism proposed in Embley et al. [in press]. Axial Seamount Province lavas are the least geochemically depleted of the four provinces discussed. They are enriched in potassium and sodium relative to other southern and central JdFR lavas (Figure 2-3), and compared to the other three aforementioned magmatic provinces, Axial Seamount lavas are less depleted in the light

REE (La/Sm_n of 0.88-0.92), Sr (Figure 2-5), $^{87}\text{Sr}/^{86}\text{Sr}$, and elemental ratios such as Zr/Y and K₂O/TiO₂. Each of these observations is indicative of geochemical enrichment in mantle-incompatible elements in Axial Seamount lavas relative to CoAxial or WFBR lavas. Lavas from Axial Seamount are not, however, obviously enriched in isotopic and incompatible elements relative to all JdFR lavas south of the Cobb offset [Rhodes et al., 1990; Smith et al., 1994]. Nonetheless, they do generally have higher values of Zr/Y, La/Sm and Sr/Zr, and, as such, the Axial data only partly overlap the geochemically enriched end of the JdFR data-field for each of these parameters [Rhodes et al., 1990]. Previous authors have modeled enrichments in Sr, Na and Ca observed in Axial Seamount lavas to be the result of a greater depth of initial melting of a source similar to that of the ridge-axis MORB [Rhodes et al., 1990]. The more depleted nature of the CoAxial Segment and WFBR provides further constraints on the degree of regional mantle geochemical heterogeneity and the depleted mantle end-member.

2.7.3.2 The Coaxial Segment

While a detailed petrogenetic analysis is beyond the intended scope of this paper, and will be dealt with elsewhere [Smith et al., in prep], a few general observations are relevant to the discussion presented here. Qualitatively, depletions in Al₂O₃ and Na₂O in the CoAxial MORB suite relative to the Axial Seamount lavas, coupled with enrichments in FeO, could be explained via increased degree of melting of an “Axial Seamount” melt source [Klein and Langmuir, 1987 and 1989, Langmuir et al., 1992]. This hypothesis is not, however, consistent with the depletion of CaO in CoAxial MORB relative to Axial Seamount lavas. In fact, depletion in CaO of CoAxial lavas relative to Axial Seamount lavas is supportive of smaller degrees of melting, unless melting proceeds well past the

elimination of CPX as a residual phase in the source mantle [Klein and Langmuir, 1987]. Additionally, the inverse correlation between Na_2O and FeO contents between the two regions corresponds to the trend associated with the “global” variability vector of Klein and Langmuir [1987, 1989], while the inverse correlation between Na_2O and SiO_2 would be more easily explained via the “local trend vector” and sampling of instantaneous melts from different depths in a melting column.

The inability to consistently account for differences in the suite of major element oxides observed between Axial Seamount and CoAxial Segment lavas via differences in the extent and depth of melting (i.e. via either the local or global trends of Klein and Langmuir, 1989) argues against the hypothesis that differences in major element characteristics between the MORB from the two regions are due to different conditions of melting of an “Axial Seamount” source in the generation CoAxial lavas.

As in the above arguments, comparison of CoAxial MORB data to the Cleft Segment MORB data also suggests that differences in major element lava chemistry between the two regions cannot be simply explained by different melting conditions of a similar and homogenous source. Lower relative Na_2O abundances in Cleft segment lavas would require a higher extent of melting relative to CoAxial lavas, but higher Al_2O_3 contents suggest a lower extent of melting. Clearly these two observations are incompatible with the above hypothesis. In summary, regional differences in major element chemistry of CoAxial lavas relative to other southern Juan de Fuca lavas cannot be generated by melting of a homogeneous mantle, but must instead be related to mantle chemical heterogeneity and or differences in crystallization and differentiation processes.

Additionally, trace element and isotopic constraints prohibit the generation of CoAxial lavas through melting of a homogeneous "Axial Seamount" source mantle.

An alternate hypothesis to that discussed above (differences in primitive melt composition relating to differences in the extent and depth of melting of similar mantle compositions) is that the source mantle sampled by Coaxial lavas is different from the other two regions either by large scale mantle heterogeneity or finer scale heterogeneity sampled differentially through variable conditions of melting. CoAxial Segment lavas exhibit characteristics of long term incompatible element depletion (radiogenically depleted $^{87}\text{Sr}/^{86}\text{Sr}$ and enriched $^{143}\text{Nd}/^{144}\text{Nd}$). Regional correlation between incompatible element ratios (Figures 2-9, 2-10 and 2-11), and between long-lived isotopes of strontium and incompatible element ratios (Figure 2-12), indicate that the range of mantle source characteristics observed can be accounted by mixing of very depleted and slightly enriched components.

Figures 2-10, 2-11 and 2-12 show regional correlations between samples from the CoAxial Segment, Axial Seamount Province, Rogue and the NRZ volcanoes, the WFBR and the Cobb-Eickelberg Seamount Chain, of which Axial Seamount is the youngest member [Desonie and Duncan, 1990]. All three diagrams show generally hyperbolic trends characteristic of mixing curves on ratio-ratio diagrams [Langmuir et al., 1978]. Figure 2-12 shows mixing of $^{87}\text{Sr}/^{86}\text{Sr}$ with both $\text{K}_2\text{O}/\text{TiO}_2$ (upper) and Sr (lower). Calculated trends are shown for mixes between a depleted CoAxial lava and lavas from Axial Seamount caldera (dashed curve) and Warwick Seamount, a member of the Cobb-Eickelberg chain (solid line). Trend symbols mark 10% mixing-increments. Examination of Figure 2-12 shows the CoAxial data trend to be elongated in the direction of the

mixing curves, with the most enriched CoAxial samples showing up to a 40% mix of “Axial Seamount melt component” or a 20% mix of a “Warwick Seamount melt component” with the depleted CoAxial end-member. Axial Seamount lavas themselves fall in the range of 40% mixing of the enriched “Warwick Seamount” end member with the depleted CoAxial end-member. While it is not the intention or within the scope of this paper to model in detail exact nature of heterogeneities occurring in the upper mantle, some interesting observations and hypothesis arise from the examination of Figures 2-10, 2-11 and 2-12.

2.7.4 Regional Mantle Heterogeneity and the Origins of CoAxial Segment Geochemical Depletion.

Regional correlation of MORB Sr isotopic values with incompatible trace element abundances implies some antiquity to mantle geochemical heterogeneity in this region. While some aspect of this heterogeneity must be related to long-term depletion events, significant variability of incompatible elements/ratios about the mixing curves at given values of $^{87}\text{Sr}/^{86}\text{Sr}$ implies there has also been more recent enrichments/depletions (Figure 2-12). This regional mantle heterogeneity is likely the result of recent (relative to the half-life of ^{87}Rb) upper mantle melting, but the cause(s) of the more ancient heterogeneity is more difficult to assess [Smith et al., in prep]. Distinct Isotopic depletions observed in CoAxial lavas relative to all other Juan de Fuca MORB (Figure 2-4) imply a depletion of ancient mantle geochemical enrichments that is unique on the Juan de Fuca Ridge. Two hypotheses are recognized to explain these depletions.

The first hypothesis is that melting around the outer portions of the Cobb melting anomaly has left “CoAxial Segment” upper mantle depleted in enriched source component. Evidence suggests that much of the northern CoAxial Segment appears

relatively amagmatic, and that the sources of recent eruptions probably have an origin south of the Floc Site. If this were in fact the case, and CoAxial lavas were being depleted by previous source melting due to the Cobb melt anomaly, then one might expect source depletions to show a radial pattern about the melt anomaly. Little evidence exists for comparably depleted lavas south of the Cobb anomaly, although sampling in this region is currently insufficient to truly test this hypothesis.

An alternate hypothesis is that recent rapid north westward migration of the ridge-axis in this area ($> 35\text{ km}$ in the past 0.5 Ma and up to 100 km in the past 7 Ma) as the ridge has approached the Cobb melting anomaly [Karsten and Delaney, 1989] has placed the current ridge-axis above a region of upper mantle already depleted through the process of MORB magma genesis. This hypothesis of rapid ridge-axis migration was called upon by Karsten and Delaney [1989] to explain the asymmetric distribution of near-axis seamounts occurring on the Pacific plate (Figure 2-1). If both the asymmetric seamount distribution and CoAxial Segment source depletions are due to the rapid northwestward migration of the ridge in this area, then one might expect two conditions to be true. Firstly, one might expect similar depletions in Vance segment just south of the melt anomaly, providing it was equally affected by ridge migration. Although existing data is sparse, this does not appear to be true. Secondly, one might expect that seamounts on the Pacific plate outside of the current ridge-axis zone of melting to have, on average, geochemical signatures more enriched than either the ridge-axis MORB or the near-axis volcanoes to their south east.

Whether the depleted character of CoAxial lavas is the result of effects associated with the Cobb melting anomaly, ridge-axis migration, or some other phenomenon, it is

likely that CoAxial lavas were produced by a mantle previously “conditioned” by at least small amounts of previous melt extraction. It is hypothesized that this previous melting preferentially sampled whatever enriched mantle component existed in the JdFR MORB source giving CoAxial lavas a distinct geochemical character relative to other JdFR lavas. Broad scale regional correlation of source characteristics and finer scale variation within individual magmatic provinces suggests that the spatial scale of mantle heterogeneity in this region is smaller than the individual magmatic provinces discussed.

2.8. Conclusions

An integrated approach to the study of MORs involving the use of geophysical, geologic and geochemical mapping methods is necessary to address, in detail, questions of spatial and temporal variability in MOR crustal accretion. At least three different eruptive events have occurred on the CoAxial Segment over a span of less than 12 years. Data presented above leads to several important conclusions regarding the process of crustal accretion at the CoAxial Segment and the spatial association of distinct melting regimes on the central JdFR in general. These conclusions include the following:

- (1) Spatial relations of these units establish a neovolcanic zone greater than 1 km wide over a decadal time scale.
- (2) Recent CoAxial Lavas are geochemically distinct from recent Axial Seamount lavas and are believed to have originated from within the CoAxial Segment, not from Axial Seamount as has been previously suggested.
- (3) Inter-flow chemical variability in CoAxial lavas suggests that the CoAxial Segment has a poorly developed magmatic plumbing system, in accord with the lack of any geophysically identified axial magma chamber within the upper 3 km of the crust.

- (4) Intra-flow geochemical variability of up to 1 wt.% MgO is observed in CoAxial lavas, and is best explained by small amounts of crystallization and differentiation along low pressure liquid lines of descent.
- (5) Four distinct magmatic provinces have been identified in the Axial Seamount-CoAxial Segment region. The range of mantle source characteristics is consistent with mixing trends established using “source” characteristics of variably depleted/enriched lavas recovered from within the region.
- (6) Correlation between long lived Sr isotopes and incompatible element ratios imply that some aspect of local mantle heterogeneity has antiquity, but scatter of incompatible element ratios at a given isotopic value further suggest that a more recent component of elemental enrichment/depletion contributes to the geochemical heterogeneity of the upper mantle in this region.

CHAPTER 3
SUBMARINE INVESTIGATIONS OF A THIRD-ORDER OSC AT 9° 37' N:
ESTABLISHING A CAUSE AND EFFECT RELATIONSHIP BETWEEN OSC
PROPOGATION AND MAGMATIC ACTIVITY

3.1. Introduction

The origins, development, and causal mechanisms of disruptions in the continuity and linearity of the East Pacific Rise (EPR) ridge-axis have been the subject of much investigation, discussion, and speculation in recent years [e.g. Macdonald and Fox, 1983 and 1988; Macdonald et al., 1984, 1986, 1987, 1988; Lonsdale, 1983 and 1986; Langmuir et al., 1986; Toomey et al., 1990; Perram and Macdonald, 1990; Haymon et al., 1991; Carbotte and Macdonald, 1992; Kent et al, 1993a, 1993b, Carbotte et al., in press, etc.]. Four different "orders" of discontinuity have been recognized on the basis of their spatial dimensions [Macdonald et al., 1988b], subdividing the northern EPR into discrete segments of varying length and temporal continuity [Macdonald and Fox, 1983; Macdonald et al., 1984; Langmuir et al., 1986; Perram and Macdonald, 1990; Toomey et al., 1990; Haymon et al., 1991; Carbotte and Macdonald 1992; Macdonald et al., 1992; etc.].

Several criteria have been used to define first through fourth-order ridge segments and associated first through fourth-order discontinuities that bound and define them [Macdonald et al., 1988b]. First-order spreading segments are tectonically defined and bounded by large transform offsets or propagating rifts whose offsets are such that plate boundaries behave rigidly, juxtaposing crust of greater than approximately 0.5-1 Myr age

difference across the offset boundary. Offsets are typically greater than 50 km and partition the ridge-axis at intervals of 100-1000 km. Second-order segments have a length scale of 50-300 km and are bounded by smaller, non-rigid discontinuities such as large overlapping spreading centers (OSC) with an offset greater than 3-5 km or small offset (less than 20 km) non-rigid transform faults. Third-order segmentation divides the ridge-axis into characteristic lengths of 30-100 km, and the segments are bounded by smaller (0.5 to 3-5 km) OSCs characterized by small (10's of meters) increases in axial depths. Lastly, fourth-order segments have typical length scales of 10-50 km. The fourth-order discontinuities are not typically associated with axial depth anomalies, and the segment boundaries are defined by small non-overlapping offsets of the ridge-axis and/or small ($1-5^\circ$) bends in the ridge-axis, referred to as deviations in axial linearity (DEVALS) [Langmuir et al., 1986]. Although first-order segments and discontinuities are thought to be persistent on the order of millions of years, smaller ridge offsets may not be as temporally stable [Macdonald et al., 1988b]. The tendency of smaller ridge-axis discontinuities to migrate and evolve has led to much speculation regarding their causative mechanism(s).

While in some instances far-field plate stresses and/or changes in the direction of spreading have been identified as the driving mechanism for migration of second and third-order discontinuities [Hey et al., 1980, 1982, 1988; Carbotte and Macdonald, 1992], magmatic activity (or lack thereof) has also been postulated as the active mechanism for the formation and evolution of similar features [Lonsdale, 1983; Sempere et al., 1984; Macdonald et al., 1984; Langmuir, 1986; Mutter et al., 1988; Toomey et al., 1990 and 1994; Smith et al., 1998; etc.]. The association between magmatic activity and fine-scale

ridge segmentation has been largely based on correlation between the location of discontinuities and factors such as axial depth, continuity and depth of the seismically imaged axial magma chamber (AMC), magnetic, seismic and structural characteristics of the crust, and the geochemistry of mid-ocean ridge basalt (MORB) dredged from either side of offsets.

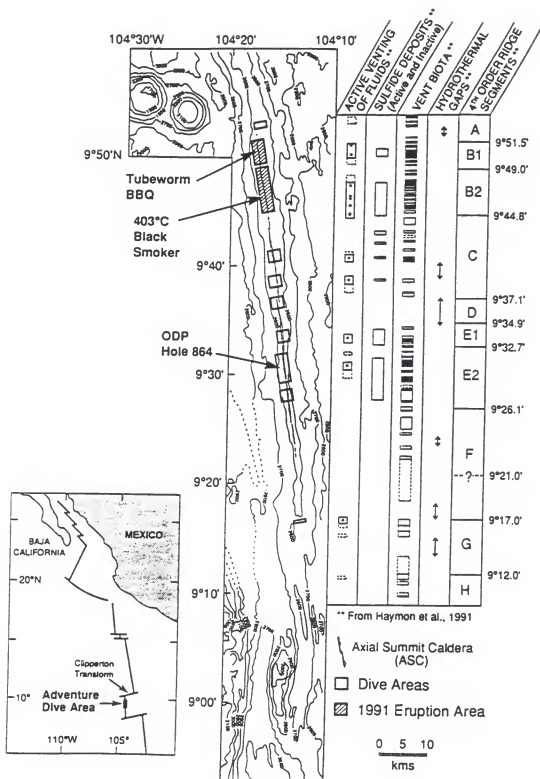
While many such discontinuities in the eastern Pacific spreading axis have been imaged bathymetrically [e.g. Macdonald et al., 1984], seismically [Toomey et al., 1990 and 1994, Kent et al., 1993a, Kent et al, 1993b; Carbotte et al, in press], and magnetically [Sempere et al., 1984], relatively few have been optically imaged or petrologically sampled [Langmuir et al., 1986; Sinton et al., 1991; Batiza and Niu, 1992]. Those that have been sampled for basalt geochemistry have been only sparsely sampled by dredge, without having the sample density or positional accuracy to reconstruct in detail the geometry of geochemical variation across the ridge-axis discontinuity. Here the first detailed report of *in situ* observational and geochemical data recovered by submersible at a small third-order discontinuity is presented. Submersible data is supplemented by dense wax core sampling of basalt glass from the crestal region surrounding the OSC.

3.2. Previous Studies and Regional Geology

3.2.1 Previous Studies

The study area lies between $\sim 9^{\circ} 35'$ and $9^{\circ} 37'$ N latitude, within a second-order spreading segment of the fast spreading EPR bounded to the south at $\sim 9^{\circ} 03'$ N by an OSC and to the north at $\sim 10^{\circ} 10'$ N by the intersection of the ridge-axis with the Clipperton transform fault (Figure 3-1) [Haymon et al., 1991]. This second-order segment spreads at approximately 11cm/yr full-rate [Klitgord and Mammerickx, 1982],

Figure 3-1: Regional bathymetric map of the second-order EPR segment between $\sim 9^{\circ} 03'$ N and 10° N latitude. Regional map-inset and scale are shown at left. Entire figure is after Haymon et al. [1991]. Contour interval for rise-crest bathymetry is 20 m and columns to right to show fourth-order segment boundaries and geologic, hydrothermal and biologic observations based on ARGO deep tow data gathered in 1989.



and has been subdivided into 10 fourth-order segments based on the occurrence of numerous small ridge-axis discontinuities identified through bathymetric, morphologic, and structural observations [Haymon et al., 1991]. No third-order axial discontinuities have been recognized along this second-order segment.

The EPR ridge-axis between 9° 09' N and 9° 54' N has been optically and acoustically imaged using the ARGO I near-bottom imaging system, with ~90% coverage over an 800 m wide swath centered on the ridge-axis [Wright et al., 1995; Haymon et al., 1991]. Subsequently, over 300 *Alvin* dives have been conducted in the region between 9° 17' N and 10° N, though greater than 80% investigated the area of 9° 46-51' N, which experienced a volcanic eruption in 1991 [Haymon et al., 1993; Gregg et al., 1996]. Since 1991, hydrothermal, geologic and biologic activity have been monitored [Fornari et al., 1998a; Von Damm, in press], and study areas have been investigated and sampled annually by submersible. Between the latitudes of 9° 17' and 10° 05' N nearly 850 MORB glass samples have been recovered by submersible and wax core, and geochemically analyzed (Figures 3-2 and 3-3) [Perfit et al., 1994 and in prep].

In addition, this segment has been well characterized through a variety of geophysical techniques including seismic reflection, refraction, and tomography [Detrick et al., 1987; Kent et al., 1993a and 1993b; Harding et al., 1993; Christeson et al., 1996 and 1997; Toomey et al., 1990 and 1994; Dunn et al., 1997; Mutter et al., 1988], seafloor compliance measurements [Crawford et al., 1999], magnetic surveys [Sempere et al., 1984; Shouten et al., 1999], and near bottom gravity surveys [Cochran et al., 1999]. Clearly, this second-order segment is among the most thoroughly studied mid-ocean ridge (MOR) environments. This study addresses both the overall character of this

Figure 3-2: Map showing regional distribution of Alvin and wax core recovered MORB. Submersible recovered samples are shown as gray filled diamonds and wax core samples as open circles. Black outline denotes the boundaries of the axial summit collapse trough (ASCT). Note the map is not a true projection, and has been adjusted to fit the page. For reference, 0.01 ° latitude or longitude is approximately equal to 1 km at this latitude.

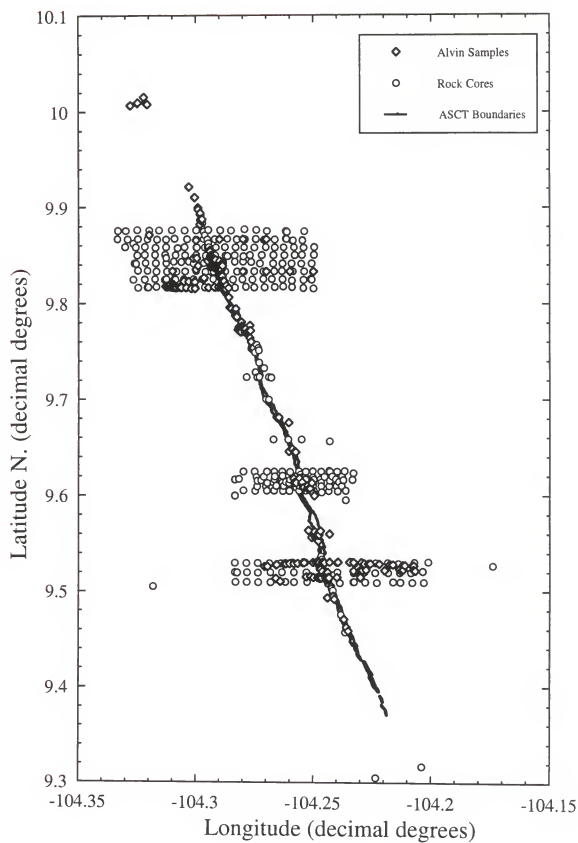
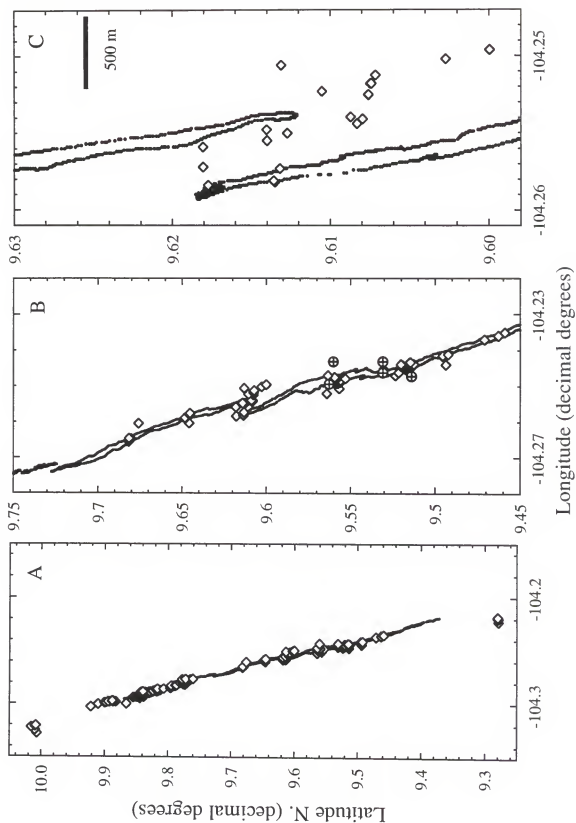


Figure 3-3: Detail maps of sample distribution for samples recovered within 300 m or less of the ASCT shown at increasingly larger scale. As in figure 3-2, maps are not true projections. Sample distributions for the entire second order segment (Panel A), segments C-E (Panel B), and the region around the 9° 37' N OSC are shown. Circles filled by crosses mark the location of mafic older samples discussed in text. ASCT boundary is identified by black outline.



second-order segment, but particularly the region between $\sim 9^{\circ} 35' - 37' \text{ N}$. A comprehensive review of existing data is well beyond the scope of this paper, and, consequently, only a brief summary of some previous findings relevant to discussions here is given below.

3.2.2 General Second-Order Scale Observations

This second-order MOR segment has a well-developed axial summit collapse trough (ASCT) [Fornari et al., 1998b] over 80% of the axis between $9^{\circ} 24'$ and $9^{\circ} 54' \text{ N}$ [Haymon et al., 1991]. Previous investigations by submersible and deep tow imaging have shown hydrothermal venting to be limited to the axial zone, and to within the ASCT when present. ASCTs are generally interpreted to form as a primary volcanic feature, but they can subsequently be modified by tectonic processes [Fornari et al., 1998b]. ASCT positions often strongly correlates with the presence of an axial magma chamber (AMC) detectable in seismic reflection surveys [Haymon et al., 1991; Carbotte et al., in press]. Ten fourth-order segments have been recognized along this portion of ridge [Haymon et al., 1991], with boundaries established by DEVALs identified at $9^{\circ} 17'$, $9^{\circ} 51.5'$, $9^{\circ} 28'$, and $9^{\circ} 35' \text{ N}$ [Langmuir et al., 1986; Toomey et al., 1990; Haymon et al., 1991] or changes in the character of the ASCT.

In general the youngest axial volcanic terrain and lava flows are found north of $9^{\circ} 37' \text{ N}$, the shallowest portion of the segment. An exception to this are young lavas that crop out near the region of the $9^{\circ} 17' \text{ N}$ DEVAL [Haymon et al., 1991; Wright et al., 1995]. Relative ages of lavas are based on data recovered within the axial zone during the 1991 ARGO survey. These data show that the youngest lavas (termed Age 1 and estimated to be on the order of less than ~ 50 years old) only crop out over three limited

regions, with the area from 9° 14'-19' being the only age 1 lavas identified south of 9° 37' N [Wright et al., 1995]. The 9° 14'-19' area is thought to have experienced an eruption sometime between 1987 and 1989 based on geologic and hydrothermal data [Fornari et al., 1998b].

North of 9° 37' N, age 1 lavas have been noted to crop out over ~90% of the length of the axis, occurring in two zones located from 9° 37.1'-42.6' N and 9° 44.3'-52' N [Wright et al., 1995]. Apparent lava ages increase southward and are generally oldest in the region between 9° 35' and 20' N. Lavas within the ASCT along this segment of ridge all have relative ages greater than Age 1 and are mostly Age 1.5–Age 2 relative to the classification scheme of Haymon et al. [1991], which is largely based on relative degrees of sediment cover and vitreous luster of lavas [Wright et al., 1995]. Although one must view the absolute ages assigned to this classification scheme with caution, lavas of age “1.5” are considered to be on the order of less than 500 years old. By comparison age “1” lavas are considered to be less than 50 years old, and age “2” lavas are assigned an age range of 1000-5000 years old. Rates of sediment accumulation and development of surficial coatings on basalt glass (thereby reducing vitreous luster) are poorly calibrated, and this type of lava dating is best suited to establishing the relative age relationships of lavas from a given region. Nonetheless, the above time scales are in general accordance with the time scales of ASCT development proposed by Fornari et al. [1998b], which were calculated based on spreading rate and ASCT widths. Additionally, they are in general accord with isotopic models of crustal residence ages for lavas from the 9°-10° N segment (Sims et al., 1997a)

Detailed examination of crustal fissuring also supports a change in the age of the ridge-axis at 9° 37' N. The 1989 ARGO survey not only revealed a general southward aging of axial lavas south of ~9° 52' N, but marked 9° 35' N as a boundary of distinct change in fissure density and average fissure width. South of 9° 35' N, fissure density increases and average fissure width decreases, both attributes associated with a predominance of tectonic activity rather than magmatic activity [Wright et al., 1995]. The southward increase in fissure density continues to the recently volcanically active region around 9° 18', where it drops again sharply. These observations are also consistent with the inferred age of the ASCT, based upon its morphological characteristics and a recently posed model for ASCT evolution [Fornari et al., 1998b].

The model of Fornari et al. [1998b] describes an evolutionary development of ASCT morphology over a time frame of approximately 5000-10,000 years whereby widening of the ASC and straightening of its bounding walls are associated with aging of the system. The ASCT is narrowest north of 9° 37' N, being only 40-80 m wide in the region from ~9° 40'-51' N. This area is also coincident with the shallowest portion of this second-order segment (~2500 m depth), and it was the site of an eruption in 1991. The eruption emanated from an 8.5 km long eruptive fissure located between 9° 46' and 51' N [Haymon et al., 1993; Rubin et al., 1994; Gregg et al., 1996]. The ASCT north of 9° 37' N are highly scalloped in plan view, and this region is considered to be in "stage 2" of the Fornari et al. (1998b) developmental cycle; perhaps 50-100 years removed from initiation of collapse trough development. The youthfulness of the crestal plateau volcanic terrain north of 9° 37' N, relative to the region further south, is supported by near bottom gravity surveys at [Cochran et al., 1999]. Cochran et al. (1999) have

proposed that because of tectonic activity and off-axis volcanism, crustal density increases with age due to collapse and infilling of voids within the extrusive layer of the crust. The surveys at 9° 50' N and 9° 30' N determined the average crustal density near 9° 50' N to be 280 kg/m³ less than that at 9° 30' N. Average crustal densities measured at 9° 30' N are typical of those determined elsewhere on the EPR and Juan de Fuca Ridge [Luyendyke, 1984; Holmes and Johnson, 1993; Stevenson 1994; Cochran et al., 1999], confirming the anomalous nature of the crustal densities around 9° 50' N.

In contrast to the region north of 9° 37' N, the area from 9° 30'-35' N contains the widest, straightest walled ASCT observed along the segment. In places, ASCT widths exceed 250 m, and this region of the ASCT has been interpreted to be in stage 4 of the conceptual ASCT developmental model. Stage 4 of ASCT development is associated with relative magmatic quiescence and widening of the ASCT due to tectonic extension [Fornari et al., 1998b]. The time scale for stage 4 ASCT development is considered to be on the order of thousands of years (but less than 10,000 years) [Fornari et al., 1998b]. ASCT morphology in this area is consistent with the observed large number of narrow, tectonically generated fissures, large cumulate fissure width and older appearance of ASCT flooring lavas [Wright et al., 1995].

3.2.3 Previous Geologic Observations of the EPR Axis Between 9° 34.9' N and 9° 37.1' N

The portion of the EPR ridge-axis between 9° 34.9' N and 9° 37.1' N has been characterized as a fourth-order segment (Segment "D" of Haymon et al., 1991) bounded to the north by an approximately 400 m right-stepping offset in the ASCT and to the south by a slight bend in the bathymetric contours of the ridge and smaller right stepping offset of the ASCT (Figure 3-1) [Haymon et al., 1991]. Closer examination of the ASCT

configuration as imaged in 1989 using the ARGO system shows there to be an overlapping relationship between the ASCTs south (western) and north (eastern) of the northern fourth-order boundary at $9^{\circ} 37.1' \text{ N}$ (Figure 3-3c). ARGO data show the eastern ASCT to be offset by approximately 0.4 km in a right-lateral sense relative to the western ASCT, overlapping it by nearly 0.75 km. Though geometrically this configuration does not correspond to the 3:1 overlap to offset ratio typical of OSCs [Macdonald et al., 1984], this feature has been referred to as a "mini OSC" [Wright et al., 1995]. This fourth-order segment (Segment "D") was identified as hydrothermal gap by Haymon et al. (1991), lacking signs of hydrothermal venting or vent biota in the 1989 ARGO survey (Figure 3-1). A more detailed analysis of the 1989 survey [Wright et al., 1995] shows there to be relatively steep gradients within Segment D in profiles of fissure width, density, and lava age versus latitude. All data indicate a transition to a more tectonically dominated regime at the southern end of this segment, when compared to the northern portion. Despite these observations, it is noteworthy that a local minima in cumulate fissure width (one of three such minima along the $9\text{-}10^{\circ}\text{N}$ segment) occurs just north of $9^{\circ} 35' \text{ N}$. Each of the other two minima on the $9\text{-}10^{\circ}\text{N}$ segment are located in regions of very recent magmatic activity ($\sim 9^{\circ} 50'$ and $9^{\circ} 18'$), and it has been proposed that local minima in the cumulate fissure width mark the areas of most recent magmatic activity along the ridge-axis.

In summary, the region of the fourth-order segment between $9^{\circ} 34.9' \text{ N}$ and $9^{\circ} 37.1' \text{ N}$ latitude marks a significant transition in geologic character of the sea floor. Differences north and south of this region are apparent in the nature and abundance of fissuring, age of axial lavas, and the morphology and developmental stage of the ASCT. Additionally, gradients in each of these characteristics can be seen within the fourth-order

segment itself, supporting the hypothesis that this portion of the ridge not only marks a boundary in the geologic character of the seafloor, but may be an area of transition between regimes of different magmatic activity. All geologic observations of the seafloor structure are consistent with the hypothesis that the ridge immediately south of Segment D is magmatically less active than the region to the north, and is instead more strongly affected by tectonic processes and amagmatic extension. As will be discussed below, petrologic and hydrologic samples recovered by submersible further support these hypotheses.

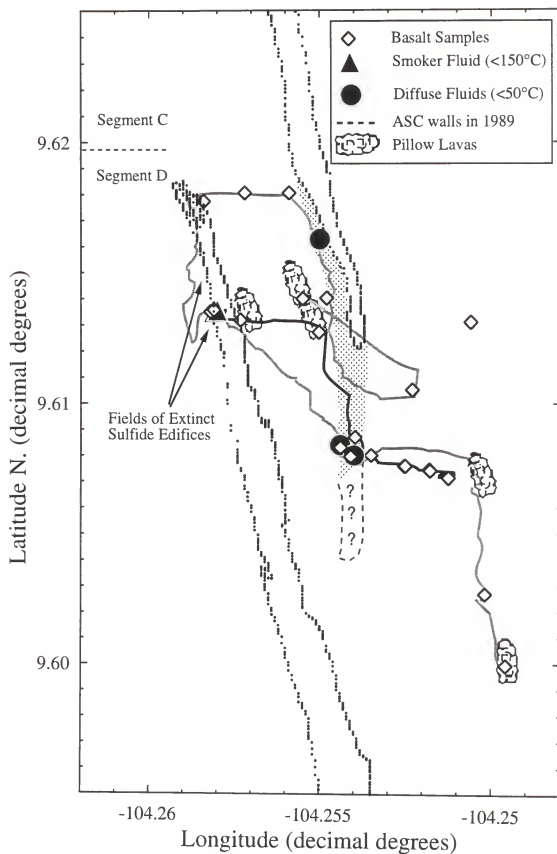
3.3. Data Acquisition and *Alvin* Observations

Detailed visual and photographic data from Segment D were obtained during *Alvin* dives in 1991 and 1994. Optical data, geologic samples and hydrothermal fluid samples were recovered primarily from two dives to this area (*Alvin* dives 2371 and 2750), and a limited amount of data was recovered during a third dive (2753) aborted early due to mechanical difficulties. Dive tracks for the completed dives and sample localities are shown in Figures 3-3c and 3-4. Dives were navigated using an array of bottom-moored acoustic transponders that provide an accuracy to within 5 m in both absolute geodetic and relative navigational frameworks [Haymon et al., 1991]. This navigational accuracy is supported by the observed ability to navigate to hydrothermal and geologic features mapped in the 1989 Argo survey to within ~10 m of their observed location [Fornari et al., 1998b].

3.3.1 *In situ* Geologic Observations.

Direct submersible observations of "Segment D" extend as far south as 9° 36' N (9.60° N). They include investigation of both the eastern and western ASCTs within the

Figure 3-4: Map showing *Alvin* dive tracks, sample locations and geologic interpretation for the region of the 9° 37' OSC. Dive track 2371 is shown as a gray line and 2750 as a black line. ASCT boundaries as mapped in 1989 are indicated by a discontinuous black outline. Basalt sample localities (gray filled diamonds), diffuse hydrothermal fluid samples (black filled circles), and higher temperature focussed hydrothermal fluids from "R" vent (black filled triangles) are shown. Stippled region represents the interpreted extent of the eastern ASCT, which both lengthened and widened between 1989 and 1991. Southern limit of ASCT advance is unconstrained by submersible observation, but based on the observation of flocculated microbial material in the water column, may extend at least as far south as the dashed line. Presence of pillow lavas is shown by graphic. Horizontal dashed line marks the segment C-D boundary after Haymon et al. [1991], and arrows mark the location of numerous extinct sulfide edifices.



zone of overlap, the region of seafloor between the two overlapping ASCTs, and the region south and east of the eastern ASCT (Figure 3-4). Dive coverage south of the overlap zone (as mapped in 1989) extends to $\sim 9^{\circ} 36.5' \text{ N}$ (9.608° N) along the extension of the eastern ASCT trend. Dive coverage continues southward from there to $\sim 9^{\circ} 36' \text{ N}$ (9.6° N), but is $\sim 300 \text{ m}$ east of that trend and 500-600 meters east of the western ASCT (Figure 3-4). Submersible coverage of Segment D's western ASCT terrain is limited to the region north of $\sim 9^{\circ} 36.7' \text{ N}$ (9.612° N). Visual and photographic data consistently show the volcanic terrain of the eastern ASCT and the southward extension of its trend to be younger in appearance than the region of the western ASCT investigated. Glassy lava surfaces of the freshest lavas in the eastern ASCT terrain (which is considered here to include a region at least 300-400 m south of its boundaries as mapped in 1989) are consistently more vitreous in appearance than the freshest lavas seen in the western ASCT.

Lavas comprising the floor of the western ASCT are largely broken up sheet and lobate lava forms, and were observed to have significant sediment cover and dull "fuzzy" appearances due to Mn and Fe oxide deposits on glass surfaces. Basalt talus piles stained bright orange by surficial coatings on the basalt were observed in numerous localities. Several extinct or weakly venting sulfide edifices were identified in this area; many of considerable size (greater than 5-8 meters in height and several meters in diameter) based on in-situ visual observations. The western ASCT extends north to approximately $9^{\circ} 37' \text{ N}$, at which point it becomes a series of discontinuous ridge parallel elongate collapse pits extending to $9^{\circ} 37.1' \text{ N}$, the northern terminus of Segment D.

In contrast to the western ASCT, the eastern ASCT, which extends from Segment "C" southward across the Segment C-D boundary has a volcanic terrain that is much younger appearance than its western counterpart. The youngest appearing lavas observed along the eastern ASCT terrain display less sediment cover than the youngest lavas of the western ASCT, and glassy lava crusts have a more vitreous appearance. The lava flows flooring the eastern ASCT are low relief lobate flows, lineated sheet flows and jumbled sheet flows. Flow contact relationships indicate that the lineated and jumbled sheet flows are stratigraphically the youngest, although the underlying lobate flows are also fresh in appearance and the contact relationship, while clear in places, is not always so.

Extensive areas of collapsed volcanic terrain indicative of ASCT development, including lava pillars and archways [Fornari et al., 1998b], extend well beyond the boundaries that were mapped in 1989 using the ARGO system [Haymon et al, 1991; Wright et al., 1995]. ASCT-like collapse extends at least as far south as $9^{\circ} 36.5' \text{ N}$, the southern most locale where the submersible survey crosses the southward extension of the eastern ASCT trend. The possibility that areas of extensive volcanogenic collapse extend even further southward along this trend cannot be precluded with the current data.

The area of seafloor in between the western and eastern ASCTs is crossed by the dive tracks in three locations, roughly in the southern, central and northern portion of submersible coverage (Figure 3-4). Old looking (Age 2) lavas and relatively heavy sediment cover generally characterize this region of seafloor. Numerous discontinuous and roughly ridge-parallel fissures in this region, and lavas forms include lobate and channelized sheet flows and several outcroppings of constructional pillow mounds or ridges (limited submersible coverage limit the ability to accurately characterize the true

dimensions of these features). The occurrence of pillow lavas in this region is surprising, as throughout the ridge between 9° 17'-52' N pillow ridges have only been observed well outside of the "neovolcanic zone" as expressed by the ASCT location. Small pillow basalt ridges/mounds were also observed eastern ridge flank some 500-600 m east of the western ASCT, and south of the southern extension of the eastern ASCT (Figure 3-4).

3.3.2 Hydrothermal and Biologic Activity

In 1991, *Alvin* dive 2371 documented extensive areas of diffuse venting of hydrothermal fluids directly out of sea floor basalt and enhanced biologic activity along the southern extension of the Segment C ASCT, as it was mapped in 1989. Diffuse venting and biologic activity extended well south of the 9° 37.1' N southern boundary of Segment C, and was confirmed as far south as 9° 36.5' N (9.61° N). No sulfide structures or focused high temperature venting was observed in this region, but diffusely venting fluids with measured temperatures of up to 39° C were observed at several sites.

Biologic activity observed in 1991 included extensive areas of seafloor blanketed by white microbial mats several cm thick. As with the area from 9° 46'-52' N affected by the 1991 eruption, this microbial biomass was pervasive and associated with abundant diffuse venting of warm hydrothermal fluids directly out of seafloor basaltic flows. In places the microbial covering of the seafloor was thick enough as to cause "white out" conditions when trying to land the submersible. In addition to the bacteria, other types of vent associated fauna were observed in the region of diffuse venting between 9° 37.1' and 36.5' N, including the tube worm species *Alvinella* and galatheid crabs. Diffuse venting of hydrothermal fluids was not only observed emanating from the seafloor, but also from the tops of lava pillars. Venting lava pillars have only been documented in two other

locations (the 9° 46-52' area and at Axial Seamount on the Juan de Fuca Ridge [Haymon et al., 1993; Perfit, personal comm.]), and in each of these cases, the area had been subject to an eruptive event in the very recent past (less than 2 years).

Diffusely venting hydrothermal fluid samples were collected in several locations along the eastern ASCT in 1991, and from a focussed, higher temperature venting sulfide structure within the western ASCT in 1991 and 1994. Data for fluids sampled from the western ASCT “black smoker” chimney, referred to as “R” vent have been reported elsewhere as part of a broader second-order segment scale investigation of hydrothermal vent fluid chemistry for this segment [Von Damm, in press]. Typically, because of the mechanics involved in the sampling of submarine hydrothermal fluids, some seawater is entrained during the sampling process because of both the “dead space” present in the sampling bottles and entrainment of ambient seawater during sampling. Not surprisingly, this effect is even more pronounced when sampling diffusely venting fluids as opposed to a more focussed flow. For this reason, fluid chemistries are typically reported as “end member” fluid chemistry by extrapolating measured values along a linear regression of their correlation with Mg, which is quantitatively removed during hydrothermal fluid-basalt interaction in the shallow ocean crust. Because pure end member hydrothermal fluids are devoid of Mg [Bischoff and Dickson, 1975], end member fluid compositions are calculated by regressing measured “impure” fluid compositions along a line that is weighted to pass through the seawater Mg value and extrapolated to $\text{Mg} = 0 \text{ mmol/kg}$.

Chloride (Cl) is the primary anionic species seawater and in hydrothermal fluids, largely because the two other major seawater anionic species, sulfate and alkalinity, are largely removed during heating of the fluid and interaction with crustal rocks. Chloride,

however, behaves in a largely conservative fashion during hydrothermal alteration of oceanic crust, and is primarily only greatly affected (in particular reduced) by the process of phase separation [Butterfield et al., 1990 and 1994; Von Damm, 1995]. As the major anion, Cl plays a large role in the flux of cationic species, which largely mobilize as Cl complexes. For that reason and because it serves as an excellent indicator of the physical process of phase separation, it is one of the most commonly measured chemical characteristics of hydrothermal fluids.

Samples of 39°C diffusely venting fluids recovered in 1991 were found to contain 519 mmol/kg of Cl (actual measured concentration) [Von Damm, unpublished data]. Extrapolation to a Mg=0 mmol/kg yields and end member hydrothermal fluid with a Cl composition of 206 ± 1 mmol/kg. This value (both the measured and extrapolated end member) is significantly less than the 540 mmol/kg of Cl present in ambient seawater indicating that these fluids represent the “vapor phase” of fluids that underwent subcritical phase separation [Von Damm, 1995 and in press]. In contrast to this, high temperature fluids ($\leq 123^\circ\text{C}$) recovered from “R” vent in 1991 and 1994 had end member Cl = 805 ± 20 mmol/kg, one and a half times that of seawater. Von Damm [in press] concluded that the chemistry of “R” vent fluids are consistent with them being the “brine phase” of phase separated hydrothermal fluids.

3.4. Basalt Geochemistry

3.4.1 Basalt Sample Recovery and Analytical Procedures.

Basalt geochemical data from Segment D include analyses of 21 samples recovered by submersible within the general axial region and 89 samples recovered by wax core from 78 sites within the broader region of the crestal plateau. Wax core

samples were recovered from an approximately 2x5 km rectangular array of sample targets centered on the OSC at 9° 37' N. Sample spacing is less than 400 m in a N-S dimension and less than 200 m in the E-W direction, and precision of sample locations is generally better than the spacing interval. The overall coverage of the sample grid is from ~9° 36.5-37.5' N latitude and 104° 14-16.8' W longitude.

Basalt glasses were analyzed for major element oxide concentration by electron microprobe (EMP) using facilities at Lamont Doherty Earth Observatory (LDEO) and the United States Geologic Survey (USGS) Branch of Geochemistry in Denver, CO. Operating conditions for EMP analyses can be found in Reynolds [1995] and Arbogast [1990] for the LDEO and USGS analyses respectively. All EMP data were beta corrected to minimize analytically generated variance due to differences in focus, polish and carbon coating. LDEO data were corrected after Reynolds [1995]. Strong correlation between sum of all oxides (Total) and an individual oxide (an indicator that the use of beta corrections are appropriate) were only observed for SiO₂ in the USGS EMP data, and therefore only SiO₂ was corrected in this data set. Measured SiO₂ values were adjusted along a linear regression between Oxide Total and SiO₂ to a Total value of 99.7, that of the MORB standard JdF. All corrected data was then normalized to the MORB standard JdF [Reynolds, 1995] which was analyzed multiple times during all analytical runs in order to facilitate direct comparison of this data to other published MORB data. Analytical precision of MORB standards run throughout the length of the analytical program are given by Rubin et al. [1998], and are similar to those presented in chapter 2. Major element data and locations for the *Alvin* and wax core samples recovered from Segment D are shown in Tables 3-1 and 3-2 respectively.

Table 3-1. Major element concentrations for Segment D *Alvin* recovered MORB Glasses Analyzed by electron microprobe (EMP)

Dive	Station	Depth	Time	Longitude	Lat. N.	#	SiO ₂	TiO ₂	Al ₂ O ₃	FeOt	MnO	MgO	CaO	Na ₂ O	K ₂ O	P ₂ O ₅	Total	K/Ti	Mg#
2371	1	2545	946	-104.2496	9.6000	7	50.2	1.54	15.1	10.0	0.20	7.9	11.8	2.69	0.13	n.d.	99.6	8.2	61.0
2371	2	2530	959	-104.2502	9.6027	5	50.4	1.58	14.7	10.3	n.d.	7.8	11.7	2.70	0.12	0.17	99.4	7.6	60.1
2371	3	2530	959	-104.2502	9.6027	5	50.6	1.25	15.0	9.5	0.20	8.4	12.0	2.56	0.07	n.d.	99.6	5.7	63.7
2371	4	2562	1056	-104.2541	9.6080	5	50.4	1.55	14.6	10.2	n.d.	7.7	11.8	2.67	0.12	0.16	99.3	7.7	60.0
2371	8	2543	1209	-104.2582	9.6135	5	50.5	1.48	15.0	9.9	n.d.	7.9	12.0	2.64	0.10	0.15	99.8	6.8	61.3
2371	9	2556	1253	-104.2584	9.6177	6	50.4	1.32	15.0	9.7	0.20	8.3	12.0	2.64	0.10	n.d.	99.7	7.6	63.0
2371	10	2555	1316	-104.2572	9.6181	5	50.4	1.76	14.3	11.0	n.d.	7.3	11.6	2.77	0.13	0.18	99.5	7.4	56.7
2371	11	2565	1329	-104.2559	9.6181	5	50.7	1.27	15.3	9.3	n.d.	8.5	12.2	2.48	0.09	0.14	99.9	7.1	64.5
2371	13	2558	1413	-104.2548	9.6140	5	50.8	1.30	15.4	9.2	n.d.	8.4	12.2	2.48	0.09	0.13	100.1	6.9	64.4
2371	14	2540	1445	-104.2523	9.6105	10	50.4	1.47	14.6	10.1	0.20	8.1	11.9	2.62	0.11	0.15	99.7	7.4	61.2
2371	15	2584	1512	-104.2555	9.6140	5	50.5	1.64	14.4	10.7	0.22	7.8	11.7	2.62	0.14	n.d.	99.6	8.3	58.8
2750	1	2527	1004	-104.2513	9.6071	4	49.7	1.49	15.7	9.6	0.21	8.2	11.7	2.78	0.13	0.16	99.8	9.0	62.8
2750	2B	2527	1026	-104.2518	9.6074	5	49.5	1.49	15.5	9.7	0.16	8.2	11.6	2.68	0.12	0.14	99.1	8.1	62.5
2750	3A	2535	1036	-104.2525	9.6076	4	50.6	1.46	14.6	9.9	0.21	8.1	11.9	2.70	0.12	0.13	99.7	8.0	61.8
2750	3B	2535	1040	-104.2525	9.6076	3	50.4	1.50	14.7	9.8	0.20	8.1	12.0	2.71	0.11	0.16	99.7	7.4	62.0
2750	4	2540	1122	-104.2540	9.6087	3	50.5	1.47	14.5	10.0	0.16	7.8	12.1	2.73	0.11	0.14	99.6	7.4	60.7
2750	6	2535	1216	-104.2550	9.6127	4	50.4	1.69	14.5	10.6	0.17	7.7	11.7	2.59	0.13	0.15	99.6	7.9	58.9
2750	7	2532	1241	-104.2573	9.6132	5	50.4	1.64	14.4	10.5	0.20	7.8	11.7	2.83	0.13	0.17	99.7	8.1	59.3
2753	1	2573	942	-104.2506	9.6131	4	50.0	1.29	15.3	9.6	0.15	8.5	12.1	2.57	0.10	0.13	99.7	7.7	63.7

notes: Dive signifies Alvin dive number; depth is in meters; Time is GMT+6 hours; # is the number of EMP analyses/sample;

All oxide concentrations reported as weight %; K/Ti is the value of K₂O/TiO₂*100, FeOt is total Fe as FeO;

Mg# is calculated as atomic Mg/(Mg+Fe²⁺) assuming 90% of Fe is FeO; n.d. - no data.

Table 3-2. Major element concentrations for MORB glasses recovered by wax core near the 9° 37' N OSC and analyzed by EMP

Sample	Depth	W/O	Longitude	Lat. N.	#	SiO ₂	TiO ₂	Al ₂ O ₃	FeO	MnO	MgO	CaO	Na ₂ O	K ₂ O	P ₂ O ₅	Total	K/Ti	Mg#
ARC-89	2585	2595	-104.2470	9.6168	6	50.3	1.53	15.3	9.9	0.19	8.0	11.6	2.67	0.11	n.d.	99.7	7.2	61.6
ARC-90	2568	2575	-104.2499	9.6167	7	50.7	1.52	15.1	9.9	0.17	7.8	11.7	2.72	0.13	n.d.	99.7	8.6	60.8
ARC-91	2556	2569	-104.2534	9.6167	5	50.3	1.30	15.2	9.4	0.18	8.5	12.3	2.56	0.12	n.d.	99.8	9.0	64.0
ARC-92	2565	2602	-104.2667	9.6168	7	50.5	1.63	15.3	9.4	0.18	7.6	11.7	2.90	0.41	n.d.	99.6	25.2	61.3
ARC-93	2257	2574	-104.2636	9.6163	7	50.4	1.58	15.3	9.4	0.17	7.7	11.7	2.93	0.40	n.d.	99.6	25.5	61.9
ARC-94	2556	2587	-104.2600	9.6164	7	50.6	1.78	14.2	11.0	0.22	7.3	11.4	2.95	0.13	n.d.	99.6	7.2	56.8
ARC-95	2565	2592	-104.2670	9.6135	6	50.7	1.63	14.9	10.0	0.18	7.5	11.7	2.81	0.26	n.d.	99.6	15.9	59.8
ARC-96	2557	2580	-104.2633	9.6134	6	50.5	1.62	15.0	9.9	0.18	7.5	11.7	2.84	0.31	n.d.	99.6	19.0	60.0
ARC-97	2557	2567	-104.2604	9.6133	6	50.6	1.77	14.2	11.1	0.19	7.3	11.5	2.85	0.12	n.d.	99.6	6.8	56.3
ARC-98	2559	2568	-104.2536	9.6135	6	50.5	1.58	14.5	10.4	0.20	7.8	11.8	2.70	0.13	n.d.	99.5	8.0	59.7
ARC-99	2556	2579	-104.2502	9.6130	7	50.1	1.28	15.3	9.3	0.19	8.6	12.1	2.60	0.10	n.d.	99.6	7.6	64.5
ARC-100	2576	2604	-104.2467	9.6137	6	50.3	1.31	15.3	9.4	0.18	8.5	12.1	2.51	0.11	n.d.	99.6	8.5	64.1
ARC-101	2555	2600	-104.2561	9.6052	7	50.2	1.66	14.3	10.9	0.19	7.6	11.6	2.92	0.14	n.d.	99.5	8.7	57.8
ARC-102	2555	2605	-104.2573	9.6084	3	50.1	1.33	15.3	9.4	0.19	8.5	12.1	2.64	0.08	n.d.	99.6	5.8	64.1
ARC-102B	2555	2605	-104.2573	9.6084	5	50.7	1.79	14.2	11.1	0.20	7.3	11.4	2.75	0.12	n.d.	99.6	6.7	56.6
ARC-103	2558	2566	-104.2554	9.6218	8	50.4	1.30	15.2	9.4	0.17	8.5	12.1	2.41	0.10	n.d.	99.6	7.5	64.3
ARC-104	2556	2589	-104.2562	9.6252	6	50.5	1.68	14.5	10.6	0.19	7.7	11.6	2.69	0.12	n.d.	99.6	6.9	59.0
ARC-121	2627	2662	-104.2334	9.6167	6	50.2	1.37	15.1	9.6	0.16	8.1	12.3	2.55	0.10	n.d.	99.6	7.2	62.5
ARC-122	2672	2684	-104.2834	9.6167	5	50.5	1.66	14.5	10.6	0.19	7.3	11.9	2.79	0.15	n.d.	99.5	8.9	57.6
ARC-200	2620	2640	-104.2433	9.6258	5	50.6	1.45	14.7	10.1	0.18	7.6	12.0	2.85	0.11	0.12	99.7	7.3	59.9
ARC-201	2604	2635	-104.2430	9.6155	5	50.2	1.56	15.0	10.4	0.20	7.6	11.7	2.67	0.13	0.15	99.7	8.3	59.3
ARC-202	2607	2646	-104.2392	9.6143	5	50.1	1.25	15.3	9.6	0.18	8.2	12.3	2.64	0.10	0.10	99.7	7.6	62.8
ARC-203	2603	2634	-104.2373	9.6163	6	50.1	1.29	15.3	9.6	0.15	8.1	12.3	2.63	0.09	0.11	99.7	7.3	62.6
ARC-218	2650	2723	-104.2817	9.6172	5	49.7	1.26	16.2	8.9	0.17	8.8	11.9	2.76	0.12	0.14	99.9	9.2	66.3
ARC-219	2600	2630	-104.2738	9.6153	5	50.6	1.62	14.3	10.6	0.20	7.7	11.8	2.65	0.10	0.15	99.7	6.2	58.9
ARC-220	2585	2622	-104.2723	9.6158	6	50.3	1.11	15.9	8.5	0.16	9.0	12.0	2.46	0.07	0.14	99.6	6.6	67.5
ARC-221	2560	2607	-104.2665	9.6200	5	50.5	1.60	14.7	10.4	0.19	7.9	11.6	2.69	0.13	0.17	99.9	8.1	59.9
ARC-222	2560	2624	-104.2605	9.6203	6	50.7	1.64	14.4	10.6	0.19	7.8	11.4	2.66	0.11	0.15	99.7	6.7	59.3
ARC-413	2614	2680	-104.2743	9.6194	5	50.7	1.32	15.0	9.3	0.17	8.1	12.5	2.65	0.11	0.13	99.8	8.6	63.4

notes: W/O stands for wire-out and is measured in meters, others as in Table 3-1

Table 3-2 continued.

Sample	Depth	W/O	Longitude	Lat. N.	#	SiO ₂	TiO ₂	Al ₂ O ₃	FeO	MnO	MgO	CaO	Na ₂ O	K ₂ O	P ₂ O ₅	Total	K/Ti	Mg#
ARC-414	2591	2630	-104.2712	9.6195	5	50.6	1.33	15.0	9.2	0.18	8.2	12.5	2.67	0.12	0.13	99.9	8.8	63.6
ARC-414A	2591	2630	-104.2712	9.6195	5	50.3	1.23	15.6	8.9	0.16	8.5	12.3	2.64	0.11	0.10	99.8	8.8	65.4
ARC-415	2565	2650	-104.2580	9.6199	5	50.2	1.53	15.4	9.3	0.18	7.8	12.1	2.81	0.38	0.21	99.9	24.6	62.5
ARC-416A	2565	2650	-104.2580	9.6199	5	50.6	1.78	14.3	11.0	0.19	7.4	11.4	2.86	0.14	0.19	99.8	7.8	57.0
ARC-416F	2565	2615	-104.2580	9.6199	5	50.4	1.81	14.3	11.0	0.20	7.5	11.7	2.83	0.15	0.19	100.0	8.2	57.4
ARC-417	2567	2596	-104.2542	9.6200	5	50.6	1.77	14.3	10.9	0.19	7.4	11.5	2.82	0.14	0.19	99.8	7.7	57.2
ARC-418	2564	2622	-104.2535	9.6199	5	50.6	1.80	14.3	11.0	0.20	7.3	11.4	2.88	0.13	0.17	99.8	7.3	57.0
ARC-419	2585	2600	-104.2502	9.6188	5	50.4	1.58	14.6	10.4	0.17	7.9	11.6	2.78	0.13	0.15	99.8	8.3	59.9
ARC-419F	2585	2600	-104.2502	9.6188	5	50.5	1.59	14.8	10.2	0.16	7.8	11.6	2.78	0.14	0.19	99.7	9.0	60.1
ARC-420	2607	2626	-104.2453	9.6196	5	50.4	1.60	14.7	10.3	0.19	7.8	11.6	2.78	0.14	0.17	99.7	8.8	59.9
ARC-420F	2607	2626	-104.2453	9.6196	5	50.5	1.62	14.9	10.3	0.19	7.7	11.6	2.78	0.13	0.17	99.8	8.0	59.6
ARC-421	2627	2665	-104.2378	9.6200	4	50.7	1.51	14.5	10.1	0.16	7.6	12.0	2.87	0.10	0.14	99.8	6.8	59.9
ARC-424	2573	2642	-104.2658	9.6103	5	50.2	1.25	15.5	8.9	0.16	8.6	12.2	2.66	0.10	0.14	99.8	7.7	65.8
ARC-425	2564	2592	-104.2627	9.6102	5	50.4	1.58	14.9	10.3	0.17	7.9	11.6	2.78	0.13	0.16	99.8	8.1	60.3
ARC-426A	2560	2570	-104.2578	9.6100	5	50.6	1.74	14.3	10.9	0.19	7.5	11.5	2.86	0.14	0.18	99.8	7.8	57.5
ARC-426F	2560	2570	-104.2578	9.6100	5	50.6	1.79	14.2	10.9	0.20	7.3	11.5	2.86	0.15	0.19	99.8	8.1	57.2
ARC-427	2563	2580	-104.2560	9.6101	5	50.6	1.41	15.0	9.7	0.16	8.1	11.9	2.68	0.10	0.14	99.8	7.4	62.4
ARC-428A	2568	2578	-104.2538	9.6098	5	50.3	1.30	15.4	9.4	0.16	8.6	12.1	2.54	0.10	0.12	99.8	8.0	64.3
ARC-428F	2568	2578	-104.2538	9.6098	5	50.5	1.45	14.9	9.7	0.18	7.9	12.1	2.69	0.11	0.16	99.8	7.6	61.5
ARC-429	2573	2594	-104.2502	9.6103	5	50.0	1.47	15.8	9.6	0.15	8.1	11.5	2.73	0.13	0.16	99.7	9.1	62.4
ARC-430	2591	2615	-104.2467	9.6100	5	50.5	1.64	15.3	9.4	0.17	7.6	11.6	2.92	0.41	0.20	99.7	24.9	61.5
ARC-431	2602	2627	-104.2433	9.6100	4	50.7	1.49	14.7	9.9	0.17	7.7	11.9	2.90	0.11	0.13	99.8	7.4	60.6
ARC-432	2606	2641	-104.2403	9.6101	4	50.3	1.60	14.9	10.2	0.17	7.8	11.7	2.77	0.13	0.13	99.7	8.0	60.3
ARC-433	2628	2643	-104.2380	9.6101	5	50.6	1.33	15.1	9.4	0.18	8.2	12.1	2.70	0.09	0.10	99.7	6.5	63.4
ARC-433A	2628	2643	-104.2380	9.6101	5	50.4	1.38	14.9	9.5	0.19	8.2	12.2	2.68	0.09	0.11	99.7	6.3	63.0
ARC-434	2670	2682	-104.2798	9.6055	4	51.0	1.55	14.2	10.3	0.18	7.7	11.7	2.90	0.12	0.14	99.7	7.7	59.5
ARC-435	2673	2695	-104.2748	9.6050	5	50.7	1.23	15.2	8.9	0.16	8.4	12.2	2.71	0.09	0.13	99.7	7.6	65.0
ARC-436	2599	2612	-104.2698	9.6051	5	50.8	1.34	14.8	9.5	0.16	8.0	12.0	2.76	0.10	0.15	99.7	7.6	62.7

notes: W/O stands for wire-out and is measured in meters, others as in Table 3-1

Table 3-2 continued.

Sample	Depth	W/O	Longitude	Lat. N.	#	SiO ₂	TiO ₂	Al ₂ O ₃	FeOt	MnO	MgO	CaO	Na ₂ O	K ₂ O	P ₂ O ₅	Total	K/Ti	Mg#
ARC-437	2582	2617	-104.2668	9.6053	5	50.9	1.60	15.1	9.3	0.16	7.7	11.4	2.91	0.40	0.18	99.6	24.8	62.0
ARC-438	2566	2591	-104.2630	9.6048	4	51.0	1.45	14.4	10.0	0.18	7.8	11.8	2.84	0.11	0.14	99.8	7.4	60.5
ARC-439	2562	2584	-104.2602	9.6049	5	50.6	1.66	14.5	10.4	0.17	7.8	11.4	2.77	0.11	0.19	99.7	6.8	59.7
ARC-440	2566	2579	-104.2533	9.6052	5	50.6	1.61	14.4	10.3	0.19	7.9	11.7	2.80	0.15	0.15	99.7	9.4	60.2
ARC-441	2564	2583	-104.2498	9.6048	5	50.3	1.64	14.8	10.5	0.19	7.6	11.7	2.79	0.11	0.14	99.8	7.0	58.9
ARC-442	2581	2593	-104.2465	9.6035	5	50.3	1.62	14.7	10.4	0.18	7.8	11.7	2.82	0.14	0.19	99.7	8.5	59.7
ARC-443	2600	2606	-104.2435	9.6049	4	50.1	1.22	15.7	8.9	0.15	8.6	12.2	2.63	0.10	0.12	99.7	8.6	65.7
ARC-444	2609	2625	-104.2397	9.6045	5	50.5	1.34	15.0	9.7	0.21	8.0	12.2	2.66	0.10	0.13	99.8	7.2	62.2
ARC-445	2629	2630	-104.2363	9.6052	5	50.8	1.52	14.4	10.4	0.22	7.5	11.9	2.84	0.12	0.16	99.8	8.1	58.9
ARC-446	2583	2600	-104.2503	9.6254	5	50.7	1.60	14.9	9.9	0.19	7.6	11.6	2.86	0.33	0.20	99.8	20.4	60.3
ARC-447	2604	2615	-104.2468	9.6251	5	50.6	1.52	14.5	10.3	0.20	7.7	11.9	2.87	0.12	0.17	99.8	8.0	59.5
ARC-448	2618	2635	-104.2430	9.6248	5	50.6	1.48	14.5	10.3	0.20	7.7	12.0	2.74	0.13	0.14	99.8	8.6	59.7
ARC-449	2640	2646	-104.2400	9.6251	5	51.1	1.62	14.2	10.9	0.24	7.2	11.3	2.94	0.14	0.15	99.8	8.9	56.6
ARC-450	2628	2642	-104.2368	9.6255	4	50.9	1.65	14.3	11.0	0.23	7.2	11.2	2.94	0.14	0.17	99.8	8.5	56.5
ARC-451	2644	2666	-104.2330	9.6234	5	50.6	1.47	14.9	9.9	0.19	7.8	11.9	2.75	0.14	0.14	99.8	9.4	60.9
ARC-452	2664	2701	-104.2798	9.6249	5	50.4	1.66	14.5	10.5	0.22	7.5	11.7	2.93	0.16	0.16	99.8	9.7	58.4
ARC-453	2602	2650	-104.2750	9.6248	5	50.4	1.63	14.5	10.7	0.23	7.6	11.6	2.77	0.14	0.17	99.8	8.9	58.5
ARC-454F	2572	2600	-104.2698	9.6250	5	50.5	1.23	15.5	8.9	0.17	8.6	12.1	2.61	0.10	0.09	99.8	7.9	65.6
ARC-454H	2572	2600	-104.2698	9.6250	5	50.3	1.22	15.5	9.0	0.20	8.6	12.2	2.63	0.11	0.11	99.8	8.7	65.5
ARC-455	2563	2592	-104.2667	9.6249	5	50.3	1.21	15.5	8.9	0.17	8.6	12.1	2.61	0.10	0.11	99.8	8.6	65.7
ARC-456H	2557	2590	-104.2633	9.6248	5	50.6	1.64	14.4	10.7	0.22	7.6	11.6	2.75	0.14	0.17	99.8	8.6	58.2
ARC-457	2555	2580	-104.2600	9.6247	5	50.6	1.74	14.2	11.0	0.22	7.4	11.5	2.82	0.15	0.15	99.8	8.7	56.9
ARC-458F	2558	2585	-104.2533	9.6249	4	50.5	1.26	15.3	9.4	0.19	8.3	12.2	2.49	0.12	0.12	100.0	9.5	63.7
ARC-516	2557	2577	-104.2632	9.6205	5	50.5	1.60	14.7	10.0	0.18	7.7	11.7	2.89	0.25	0.16	99.7	15.5	60.3
ARC-517	2613	2638	-104.2420	9.6199	5	50.4	1.64	14.4	10.4	0.20	7.6	11.8	2.88	0.15	0.17	99.7	8.9	59.0
ARC-518H	2600	2636	-104.2405	9.6171	5	50.0	1.35	15.1	9.6	0.16	8.3	12.2	2.68	0.08	0.14	99.6	6.3	63.2
ARC-519	2556	2575	-104.2617	9.6142	5	50.2	1.81	14.2	10.9	0.19	7.5	11.5	2.86	0.14	0.14	99.5	7.5	57.4
ARC-520	2583	2594	-104.2697	9.6131	5	50.6	1.64	14.9	10.0	0.17	7.6	11.7	2.88	0.30	0.20	99.9	18.6	59.9

notes: W/O stands for wire-out and is measured in meters, others as in Table 3-1

Table 3-3. Trace element concentration in representative samples analyzed by laser ablation inductively coupled mass spectrometry (LA ICP-MS)

Element	P. Mantle	2495-5	2372-1	ARC 93	ARC 95	ARC 97	ARC 99
Cs	0.032	0.08	0.012	0.101	0.038	0.013	0.007
Rb	0.635	6.1	0.65	4.1	2.4	0.84	0.46
Ba	6.989	36	8.5	47	28	9.2	6.3
Th	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.
U	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.
Nb	0.713	9.8	2.1	8.8	5.7	2.5	1.6
Ta	0.041	<0.2	0.19	0.7	0.46	0.2	0.13
K ₂ O	0.0301	0.36	0.09	0.42	0.26	0.13	0.10
La	0.687	9.2	3.7	8.2	6.0	4.3	3.0
Ce	1.775	22.0	10.8	18.9	14.8	12.6	8.8
Pr	0.276	3.3	1.9	2.9	2.3	2.3	1.5
Sr	21	165	126	183	147	117	114
Nd	1.4	17.0	10.2	13.9	11.9	12.1	8.2
Zr	11.2	151	111	139	128	134	89
Hf	0.309	3.8	3.0	3.6	3.3	3.7	2.5
Sm	0.444	4.6	3.6	4.2	3.8	4.3	2.9
Eu	0.168	1.6	1.4	1.6	1.4	1.5	1.1
TiO ₂	0.21685	1.8	1.6	1.6	1.7	1.8	1.3
Tb	0.108	1	0.93	0.97	0.96	1.11	0.76
Dy	0.737	6.4	6.4	6.6	6.5	7.4	5.4
Y	4.55	39	35	35	35	41	29
Ho	0.164	1.4	1.3	1.3	1.3	1.6	1.1
Er	0.48	3.9	3.9	3.8	3.8	4.6	3.2
Yb	0.493	3.7	3.7	3.6	3.7	4.2	3.1

notes: All data reported as parts per million (ppm) except K₂O and TiO₂ which are reported as weight % oxide. Primitive mantle concentrations after Sun and McDonough [1989]

A limited subset of wax core samples was analyzed for trace element concentrations by laser ablation inductively coupled plasma source mass spectrometry (LA ICP-MS) at the USGS in Denver. Details of operational procedures, instrumental conditions, and analytical detection limits and precision are given by Ridley and Lichte [1998]. Concentrations of select trace elements for samples analyzed by LA ICP-MS are shown in Table 3-3.

Examination of MORB major element data from the entire region between $\sim 9^{\circ}$ 17' N and 10° N (Perfit et al., 1994; our unpublished data) and comparison of that data to the data from Segment D basalt samples reveals several notable relationships.

3.4.2 Second-Order Segment Scale Geochemical Observations

Nearly 850 well located basalt samples from the EPR axis and crestal plateau between $\sim 9^{\circ}$ 17' and 10° N have been analyzed for element concentrations [Perfit et al., in prep.]. The majority of these samples have elemental compositions characteristic of normal incompatible element depleted mid-ocean ridge basalt (N-MORB). A small, but significant, percentage (less than $\sim 15\%$) of the samples recovered outside of the ridge-axis on the crestal plateau have elemental signatures that are transitional to incompatible element enriched basalt. These elementally enriched MORB have been referred to as T-MORB or E-MORB (referring to their transitional and enriched character), depending on the level of incompatible element enrichments observed [Sinton et al., 1991; Batiza and Niu, 1992; Reynolds et al., 1992; Perfit et al., 1994]. The boundary at which a T-MORB becomes an E-MORB has not been well defined, and this group of basalts has generally been treated as a continuum of compositions resulting from the mixing of enriched and depleted melts or melt sources. As such, this paper will simply refer to the group of

elementally enriched MORB recovered on the 9-10° N segment as E-MORB, noting that this designation may not correspond to similar levels of elemental enrichment reserved for the E-MORB classification in other existing literature.

Differences in the ratios of highly incompatible elements [Perfit et al. 1994; Batiza and Niu, 1992] and isotopic signatures [Sims et al., 1997b] have led investigators to conclude that E-MORB from the 9-10° N segment of the EPR (and E-MORB in general) do not have a direct genetic link to the more voluminous N-MORB. It has been further hypothesized that these E-MORB have retained their geochemically enriched signatures by avoiding significant mixing in an axial magma chamber [Reynolds et al., 1992; Perfit et al., 1994; Perfit and Chadwick, 1998]. This conclusion is consistent with the observation that no E-MORB have been recovered within or proximal to the ASCT between 9° 17' and 10° N where an AMC has been imaged [Perfit and Chadwick, 1998]. In order to accurately assess recent axial magmatic conditions and processes, the geochemical database for the 9-10° N segment requires filtering to remove data from E-MORB, which are considered to have origins distinct from typical AMC melts. Because many of the samples in question were recovered by wax core and have only been analyzed for major element concentrations, the ratio K_2O/TiO_2 serves as the best available discriminator.

Examination of the relationship between K_2O/TiO_2 and MgO measured in MORB glasses (Figure 3-5) reveals several important observations allowing for the discrimination of E-MORB. Firstly, no significant correlation exists between K_2O/TiO_2 and MgO. Because MgO is strongly controlled by crystallization and assimilation processes, a lack of correlation with K_2O/TiO_2 implies that K_2O/TiO_2 values are

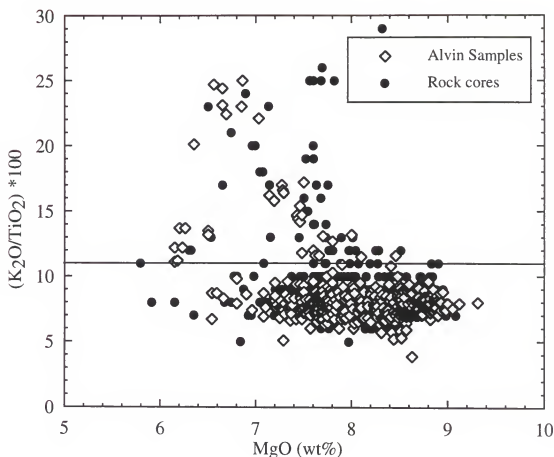


Figure 3-5: K_2O/TiO_2 plotted against MgO (weight %) for all lavas recovered from the 9°-10° N second order segment. Maximum possible variation in K_2O/TiO_2 is less than 16% (c.f. Figure 2-9) of the value. No significant correlation between MgO and K_2O/TiO_2 is observed. E-MORB fall above the horizontal line at a K_2O/TiO_2 value of 11 or greater (see text).

established during magma genesis in the mantle, and that subsequent crustal level modification would most likely be the result of mixing processes. That K_2O/TiO_2 is a genetic trait inherited from parental magmas confirms its use as a discriminator of EPR magmas with different genetic origins [Langmuir et al., 1986; Sinton et al., 1991; Reynolds et al., 1992; Perfit et al., 1994]. Additionally, investigations of other ridge segments have also shown K_2O/TiO_2 to be an accurate measure of mantle derived geochemical enrichment by showing strong correlation between K_2O/TiO_2 and other mantle derived geochemical characteristics including Sr isotopic ratio, Zr/Y and La/Sm [Smith et al., in prep].

Secondly, examination of Figure 3-5 reveals a break in the majority of the K_2O/TiO_2 data at a value of $K_2O/TiO_2 = 11$, with the vast majority of samples having a value of ≤ 10 (rounded to the nearest whole number which is appropriate to the analytical precision of this parameter). That this K_2O/TiO_2 level serves as a reasonable dividing line between the N- and E-MORB is supported by several observations. Firstly, if one evaluates the distribution of K_2O/TiO_2 values observed in MORB recovered from the wax core grids that sampled the crestal plateau regions around $9^\circ 30' N$ and $9^\circ 37' N$ (Figure 3-2), one finds that there is a bimodal distribution of values with K_2O/TiO_2 being either ≤ 10 or ≥ 15 . No values between 10 and 15 are in evidence in what is otherwise a relative continuum of values between ~ 5 and 25. This relationship, however, is not as clear in data from the wax core grid centered around $9^\circ 50' N$ where a limited number of samples recovered outside of the axial region have K_2O/TiO_2 intermediate between 10 and 15 ($\sim 15\%$ of 189 samples recovered from the sampling grid). Secondly, only 1 of 294 analyzed samples recovered within or proximal to (within 300-500 m) the ASCT (i.e.

directly associated with the axial melt lens) has a K_2O/TiO_2 value greater than 10, and that one sample is an age 2 lava from the $9^\circ 50'$ N region, recovered outside of the ASCT. This observation implies that lavas along the 9 - 10° N segment, lavas emanating from the axial melt lens characteristically have $K_2O/TiO_2 \leq 10$.

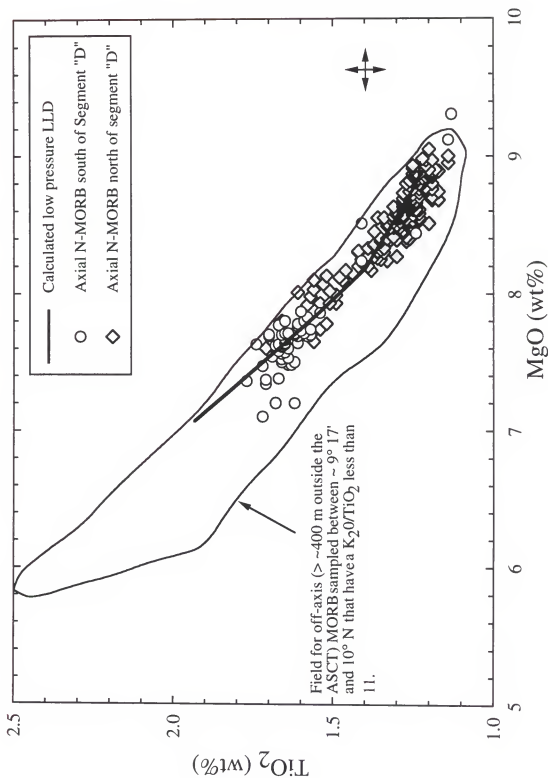
Lastly, an examination of the “off-axis” crestal plateau lavas in the $9^\circ 50'$ region, for which there is greater submersible coverage and a greater amount of trace element data that can be correlated to the major element data, shows there to be mantle derived trace element distinctions between lavas with $K_2O/TiO_2 \leq 10$ and those with values greater than 10. These distinctions include enrichments in both Sr abundance and the ratio Zr/Y in samples with K_2O/TiO_2 greater than 10 [Perfit et al., in prep.]. For all of the above reasons, samples w/ a $K_2O/TiO_2 \geq 11$ have been filtered out of the N-MORB database, being considered E-MORB. This value is roughly similar to the value of 9 used to discriminate N-MORB from T-MORB on the EPR at $\sim 12^\circ$ N [Reynolds et al., 1992]. After exclusion of the aforementioned E-MORB, 695 analyzed basalt samples recovered on- and off-axis by submersible (378) or wax core (317) are included in the N-MORB database.

In order to discriminate the most recent axial magmatic activity from possible longer temporal signals, further subdivision of the 9° - 10° N N-MORB database is required, distinguishing those samples recovered very proximal to (less than ~ 300 m) or within the ASCT from those recovered further off-axis. Although evidence suggests that eruptive activity (including both N- and E-MORB) can occur on this segment as far as 4 km off-axis [Perfit et al., 1994], the inability to unambiguously evaluate the entire database as to which of the off-axis samples recovered were erupted within the axis then

moved off-axis by subsequent seafloor spreading and which were more recently erupted off-axis requires that those samples considered to be most recent include only samples recovered either within or proximal to the ASCT (within a ~300m). After applying this second filter to the submersible recovered database (very few wax cores were recovered from within the axis) 272 axial N-MORB analyses remain (192 north of Segment D, 58 south of Segment D and 22 within Segment D).

Axial N-MORB data for the oxides MgO and TiO₂ are shown in Figure 3-6. A field that encompasses the total range of all off-axis N-MORB sampled between 9° 17' and 10° N is shown for reference. Axial MORB exhibit an overall range of MgO content from ~9.3 to 7.1 wt.%. Consideration of off-axis N-MORB extends the range to more evolved (relative to primitive, high MgO MORB) compositions with MgO as low as 5.9 weight percent (wt.%). The overall trend of data for axial N-MORB (and off-axis N-MORB as well) corresponds well to a calculated liquid line of descent (LLD) that assumes low-pressure (1kbar) fractional crystallization as the magmatic differentiation process [Weaver and Langmuir, 1990]. This implies that the dominant control of geochemical heterogeneity among the N-MORB suite in the 9°-10°N region is cooling, crystallization and subsequent differentiation of magma bodies within the shallow oceanic lithosphere. This conclusion is in accord with many other investigations of medium to fast spreading ridge environments [e.g. Batiza and Niu, 1992; Smith et al., 1994]. While significant heterogeneity in MgO content within this lava suite exists, compositional variation at a given MgO is quite limited. For N-MORB recovered from within the axis, variation in TiO₂ at a given MgO content is representative of the variation observed in the other major oxides at a given MgO and is quite limited, being

Figure 3-6: Covariation of TiO_2 and MgO for Alvin recovered axial MORB. Field denotes compositional range for all off-axis N-Type MORB ($n=392$) recovered by rock core ($n=317$) or submersible ($n=75$). Symbols are as in legend. Arrows indicating ± 2 -sigma errors are shown in lower right and indicate $\pm 2\%$ of MgO and 3% of TiO_2 for the most evolved axial basalt. A low-pressure (1 kilobar) liquid line of decent is shown for fractional crystallization of a mafic basalt sampled from the 1991 eruption near $9^\circ 50' \text{ N}$ latitude.



barely outside of (and less than twice that of) analytical uncertainty. This leads to the conclusion that the entire range of observed compositions in axial N-MORB can be generated by similar crystallization conditions of very similar parent magma compositions. That axial basalt in this area can all be generated from common or very similar parental compositions was also noted by Batiza et al. [1992], who, unlike Langmuir et al. [1986] found no evidence of major changes on MORB parental chemistry associated with fourth-order ridge-axis discontinuities in this region. Inclusion of off-axis samples does, however, as much as double the range of TiO_2 observed at a given value of MgO .

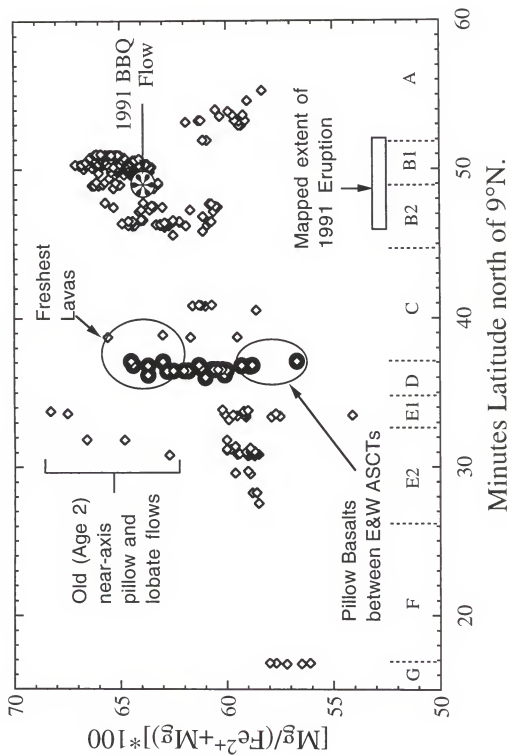
In order to facilitate the evaluation that Segment D marks a significant change in the character of the ridge crest along the 9° - 10°N segment, axial MORB samples plotted in Figure 3-6 have been divided into two subsets, those south of Segment D (open circles) and those north of Segment D (gray filled diamonds). Note that samples from the axis of Segment D itself are not shown on this diagram, but fall within the overall field for axial N-MORB. Samples from each of these subsets fall into distinctly different groups based upon their MgO content (and consequently many other elements displaying strong correlation to MgO), which is the best available discriminator since all axial N-MORB along this second-order segment appear to be derived from similar parents and crustal differentiation processes.

North of Segment D, 95% of all (192) axial samples have a MgO content greater than 7.9 wt.%, with none having compositions more evolved than 7.6 wt.% MgO . All of the samples north of Segment D that are more evolved than 7.9 wt.% MgO were recovered on Segment C, the fourth-order segment adjacent to, and immediately north of

Segment D. South of Segment D, the majority of axial lavas fall within a more limited compositional range of 7.1-7.9 wt.% MgO, with only 5 of 58 samples falling outside of (to higher MgO contents) that range. Liquid line of descent calculations [Weaver and Langmuir, 1990] give melt temperature estimates that ranging from 1210-1170°C for the axial N-MORB glass data shown in Figure 3-6. With the exception of the five aforementioned samples, data south of Segment D have calculated melt temperatures less than 1185°C whereas those north of Segment D generally have calculated melt temperatures greater than 1185°C.

Figure 3-7 illustrates chemical variation along the second-order segment by examining the relationship between latitude and Magnesium number (Mg#). Mg# represents the ratio of atomic Mg to the sum of atomic Mg and Fe^{2+} , and it is strongly correlated to both MgO abundance and magma temperature. If one excludes the Segment D data which exhibits a broader range of Mg# values than at any other latitude on the 9°-10° N segment, a strong positive correlation exists between latitude and Mg# for the portion of ridge south of Segment A. An exception to this are the five aforementioned samples south of Segment D that have compositions anomalously mafic (MgO rich) for their latitude. These 5 samples are indicated in Figure 3-3b by circles filled by crosses, and are consistently among the oldest (all age 2) samples identified in the region of their recovery. Linear regression of the remaining data yields a r^2 value of 0.807. Smooth latitudinal correlation of Mg# at the second-order segment scale have previously been noted for this segment and portions of the southern EPR [Batiza and Niu, 1992; Sinton et al., 1991].

Figure 3-7: Correlation between lava Mg# ($\text{Mg}/(\text{Mg}+\text{Fe}^{2+})$) and latitude (minutes north of 9° N) for axial N-MORB (gray filled diamonds) between 9° 17' N and 10° N. Samples recovered by submersible in the vicinity of the 9° 37' OSC are shown by filled circles surrounding the diamonds. Mg # correlates strongly with both MgO content and melt temperature. The 9° 37' OSC shows the widest range of Mg#, and therefore melt temperature, at a given latitude of anywhere on the ridge axis. The youngest lavas (samples within the circle) have Mg# anomalously high relative to the regional trend, and lavas between the two OSC limbs have Mg#s that are anomalously low relative to the regional trend. Anomalously high Mg# lavas south of segment "D" discussed in text are indicated by text and brackets. Fourth-order segment boundaries after Haymon et al. [1991] are marked at the bottom by vertical dashed lines. The gray bar marks the area over which the 1991 eruptive has been confirmed to outcrop [Gregg et al., 1996], and the circled asterisk the average composition of that flow [Perfit and Chadwick, 1998].



3.4.3 Fourth-Order Segment Scale Geochemical Observations of Segment D

Some limited, but significant, systematic relationships between lava chemistry and ridge-axis (ASCT) geometry are observed in the region of the 9° 37' N OSC. The youngest lava flows flooring the eastern limb ASCT (which extends southward from Segment C) are the most mafic recovered along the segment having MgO concentrations of 8.4-8.5 wt.%, and they fall well above the regional trend in Mg# established in Figure 3-7. These young sheet flows (samples 2371-3, 11 and 13) were sampled in three localities along the eastern ASCT over an approximately 1 km length of ridge, and all have major element compositions indistinguishable from one another. Additionally, these lavas have Mg#s ranging from 63.7 to 64.5 (indistinguishable given analytical uncertainty) identical to the average Mg# of 64 measured in the 1991 eruptive to the north suggesting they may be related to the same pulse of magmatic/volcanic activity. Other very fresh (age 1), but slightly older lavas from the eastern ASTC and slightly east of it (2371-4, 5, 14 and 2750-3a, 3b and 4) tend to have lobate and sheet forms, and have slightly more evolved compositions ranging between 7.8 and 8.1 wt.% MgO.

MORB recovered from the ASCT associated with the western OSC limb are limited to two samples, but the few sampled are more chemically evolved than the most recent flows on the eastern ASCT. The most evolved (lowest MgO content) lavas are consistently found in the region of overlap between the two ASCTs. In this region pillow lavas are common and the four samples recovered from this region (2371-10 and 15 and 2750-6 and 7) all have MgO contents less than 7.8 wt.% (as low as 7.3) and Mg#s below the regional trend.

3.5. Discussion

3.5.1 Southward Aging of Volcanic Activity

Numerous geologic observations support the hypothesis that there is an overall southward decrease in magmatic input and volcanic activity between $\sim 9^{\circ} 52'$ and $9^{\circ} 17'$ N. Differences in axial depth, density and character of crustal fissuring, size and shape of the ASCT, apparent age of axial lavas, crustal density, and lava eruption temperature (Mg#) all are suggestive that southward along the ridge-axis conditions become less magmatically/volcanically robust, the axis has a more mature volcanic terrain, and it is more affected by tectonic stretching and amagmatic extension [Haymon et al., 1991; Wright et al., 1995; Fornari et al., 1998b; Cochran et al., 1999]. The southward decrease in melt temperature (Mg#) illustrated in Figure 3-7 has also been previously recognized by Batiza and Niu [1992], although it is not as well constrained in their data as it is by data shown in Figure 3-7. The poorer resolution and greater scatter about the trend presented by Batiza and Niu [1992] is most likely due to the lower sampling density (dredges $\sim$ every 1.8 km), lesser overall number of samples involved in the study, lack of sample recovery north of $9^{\circ} 40.5'$ N, and the poorer spatial control in dredging as compared to recovery by submersible or wax core. Poorer precision in the location of sample recovery limits the ability to efficiently filter out true axial lavas from those that were near, but outside of the zone dominated by recent volcanics erupted from the axial melt lens.

Rather than calling for differences in magmatic input to the AMC to account for observed differences in melt temperature, Batiza and Niu [1992] explain the observed southward decrease in melt temperature by postulating the presence of a continuous,

chemically zoned magma chamber over the entire second-order segment. According to their model, melt is centrally supplied and cooler, denser, Fe-rich melts gravitationally migrate to the distal, deeper ends of the axial magma chamber. This model makes the assumption that the numerous fourth-order discontinuities present over this second-order segment do not correspond to discontinuities within the AMC that would inhibit such along axis crustal magma migration. This assumption will be further addressed below.

Inclusion of N-MORB data from the three wax core grids (Figure 3-2) with the axial data presented in Figure 3-7 greatly reduces, if not completely eliminates the latitudinal correlation with Mg#. This observation may have important ramifications for the time scale of variability in magmatic input to the ridge. The wax core grids at $\sim 9^{\circ} 30'$, $9^{\circ} 37'$, and $9^{\circ} 50'$ N extend up to 4 km on either side of the ridge-axis. At a full spreading rate of 11 cm/yr, this area represents in excess of 70,000 years of seafloor spreading. On the one hand, correlation of lava geochemistry and latitude seen in ridge-axis lavas, but not observed over the broader crestal region may reflect a sudden and recent change in the behavior of this portion of the ridge-axis. This hypothesis seems unjustified and contrary to the principle of uniformitarianism.

If, on the other hand, ridge-axis magmatism waxes and wanes through time as suggested by other segment-scale studies of intermediate- and fast-spreading ridges [Reynolds et al., 1992; Smith et al., 1994; Batiza et al, 1996], and the current observations of axial zone geology and MORB geochemistry are therefore a snapshot in the temporal evolution of this waxing and waning, then this cycle must occur on a time scale finer than the 70,000 years of crustal accretion represented in the area of wax core survey. Thus, the spatial signal would be muted by the inclusion of several such

temporal cycles within the areal/temporal extent of the wax core surveys. This time scale of temporal magmatic evolution (on the order of 10^3 - 10^4 years) is generally consistent with estimations of the range of lava ages observed within the neovolcanic zone of the segment and consistent with time scales of formation for observed variations in ASCT morphology. It is, however, important to note that volcanism and crustal accretion are not limited to occurring within the axis.

Though its volumetric significance is as of yet to be established, Perfit et al. [1994] established that recent volcanic activity has occurred up to 4 km outside the ASCT near $9^{\circ} 31.5' \text{ N}$, greatly confusing the relationship between distance from the ridge-axis and lava age. Several authors have recognized that on the EPR the seismic layer 2a (thought to represent the extrusive portion of the crust) increases in thickness by up to a factor of two within a few km of the axis [Harding et al., 1993; Christeson et al., 1994; Vera and Diebold, 1994]. Some have proposed periodic volcanic flooding of the crestral plateau as a mechanism to accomplish the thickening [Hooft et al., 1996]. Given these observations, it is likely that off-axis volcanism will at least dampen any temporal geochemical signal established in the ridge-axis, and it possibly may dominate the broader crestral plateau data. Exactly what the magnitude of off-axis volcanism's effect on the overall geochemical character of the crestral plateau cannot be adequately addressed until a more precise means of distinguishing axial from off-axis volcanics is achieved.

3.5.2 The $9^{\circ} 37' \text{ N}$ OSC as a Magmatic and Hydrothermal Divide

Several observations support that the $9^{\circ} 37' \text{ N}$ OSC marks a boundary between different magmatic and hydrothermal systems. Conditions of crustal structure such as

lava age, fissure density and fissure width make an abrupt rather than continuous change in character at Segment D [Wright et al, 1995]. The large difference in the morphology and apparent age between the eastern and western ASCTs of the OSC, as well as the apparent difference in the age of lavas flooring them, strongly imply that the effects of discrete eruptive events do not extend across the OSC. This contention is further supported by significant differences in hydrothermal venting across the OSC. Numerous large sulfide edifices, both venting and extinct, have been identified within the ASCT of the western limb of the OSC, while no sulfide structures or high temperature focussed hydrothermal venting have been observed within the ASCT of the eastern limb. Vent fluid chemistries north of the OSC have chemistries very different from those south of the OSC.

Von Damm [in press] noted that in 1991 hydrothermal vents north of the 9° 37' N OSC (Segments B and C) were commonly venting fluids in excess of 350°C, and that all vents north of that boundary vented hydrothermal fluids with end member chlorinity less than that of seawater. Von Damm [in press] also noted that those hydrothermal vents south of the OSC and north of 9° 17' N (including "R" vent which is in the ASCT of the western OSC limb) all vent hydrothermal fluids with temperatures generally below 300° and end member chlorinity greater than that of seawater. End member hydrothermal fluids with chlorinity less than that of seawater represent the vapor phase of fluids that have undergone the process of subcritical phase separation. Low end-member fluid chlorinities, previously unprecedented rapid changes in fluid chemistry in individual vents, shallow depth of fluid reaction (as evidenced by low end member Si contents), and short apparent crustal residence times of fluids (as evidenced by low Li/Cl and K/Cl

ratios) led to the conclusion that vents north of the 9° 37' OSC had been affected by the 1991 volcanic eruption between 9° 46 and 52' N [Von Damm, in press]. Conversely, no vents south of the OSC appear to have been effected by the 1991 event. In fact, vents south of the OSC (but north of the recently volcanically active 9° 17' N area) have end member fluid chemistry consistent with being the brine phase of phase separated fluids. If phase separation of hydrothermal fluids is a common process associated with seafloor volcanism, and vapor phase fluids are vented after eruption, venting of the conjugate brine phase may indicate that it has been some time since the hydrothermal system was volcanically perturbed [Von Damm, personal comm.].

All of the above geologic, petrologic and hydrologic arguments above strongly suggest that the 9° 37' N OSC marks a barrier to communication between the magmatic and hydrothermal systems north and south of it. This hypothesis is further bolstered by seismic studies of this region.

Seismic tomography and 2-d seismic reflection surveys of the region around the 9° 37' N OSC provide information independent of petrologic and hydrothermal data that is relevant to magmatic activity. Kent et al. [1993b] used seismic reflection common depth point profiles to image the AMC and distribution of magma along this portion of the segment. Although they found the AMC to be continuous across the 9° 17' and 35' DEVALS, Kent et al. [1993b] concluded that the DEVALS each mark a place where the melt lens exhibits a right lateral offset and melt connectivity may be disrupted. Additionally, the 9° 35' DEVAL is associated with the narrowest AMC (250 m) observed on the 9-10° segment. These conclusions are consistent with tomographic data indicating that DEVALS at 9° 28' and 35' mark perturbations in crustal the crustal velocity

structure of the ridge-axis [Toomey et al., 1990 and 1994]. Maximum low-velocity anomalies at the center of the fourth-order segments that taper off towards the segment ends (at the DEVALS) led Toomey et al. [1990 and 1994] and Dunn and Toomey [1997] to conclude that along-axis segmentation resulted from melt injection at fourth-order segment midpoints (at ~10 km intervals). DEVALS were hypothesized to mark boundaries between different melt injection regimes. The above geophysical arguments support the hypothesis that the 9° 37' OSC marks a divide between different crustal magmatic plumbing systems.

3.5.3 Evidence for a magmatic perturbation at 9° 37'

3.5.3.1 Hydrothermal and biologic evidence for recent volcanic activity

Many observations support the hypothesis that the eastern limb of the 9° 37' OSC was affected by a magmatic perturbation between 1989 and 1991. The abundant microbial material covering the seafloor in this area in 1991 was not previously observed in the 1989 ARGO surveys, nor was it present afterward in 1994. An abundance of microbial material covering the seafloor has been only previously observed associated with very recent volcanic activity [Haymon et al., 1993; Embley et al, in press]. Based on time-series investigations of recent eruptive sites on the Juan de Fuca Ridge (Cleft and CoAxial Segments) and the East Pacific Rise (~9° 46-52' N) these bacterial blooms appear to be a relatively short-lived phenomenon. Abundant diffuse venting of hydrothermal fluids directly out of the seafloor as observed on the eastern OSC limb, is also thought to be a relatively short lived condition occurring in areas that have been recently volcanically active [Haymon et al., 1993; Embley et al., in press]. More than just the occurrence of abundant diffuse venting, the chemistry of the vent fluids

themselves strongly indicate a very recent magmatic perturbation. Low end member chlorinity of diffuse fluids sampled along the eastern OSC limb indicate venting of vapor phase fluids similar to vents on segments B and C affected by the 1991 eruption at 9° 50'. Additionally, these diffuse fluids have anomalously high-end member ratio of $^3\text{He}/\text{heat}$ [Lupton, unpublished data]. Anomalously high $^3\text{He}/\text{heat}$ is also associated with venting shortly after a volcanic eruption [Baker and Lupton, 1990]. Unlike the phase-separated fluids, high $^3\text{He}/\text{heat}$ requires not just a thermal input to the system, but a magmatic input to supply the excess ^3He .

3.5.3.2 Southward propagation of the eastern OSC limb

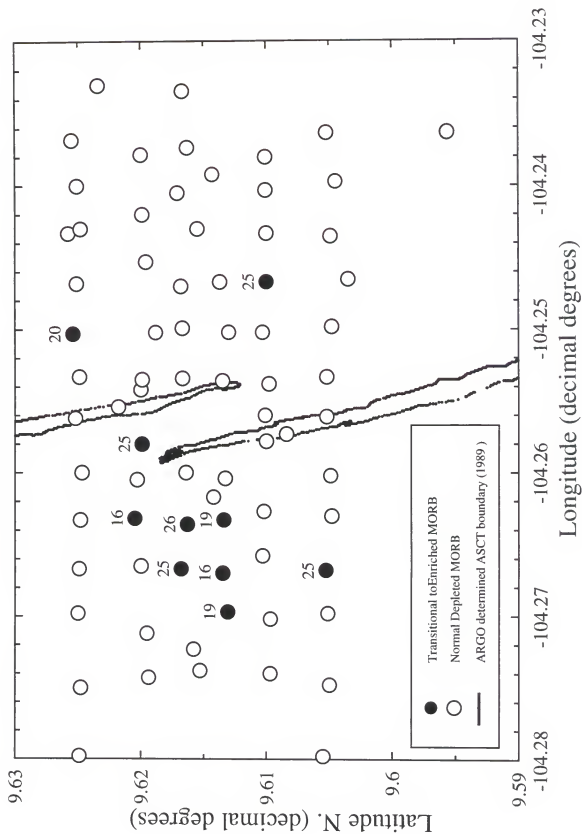
In addition to biological and hydrothermal evidence, changes in the areal extent of the eastern limb ASCT between 1989 and 1991 also suggest a volcanic event occurred during this time frame. Figure 3-4 illustrates changes in the ASCT configuration between 1989 and 1991. Interpretations are based on submersible observations and indicate that the ASCT extended southward by more than 400 m and widened by several 10's of m between 1989 and 1991. This extension of the ASCT changes the overlap to offset ratio to ~ 3 , the dimensional ratio typical for overlapping spreading centers along the EPR [Macdonald et al., 1984]. It is also intriguing that very fresh lavas, with a similar chemistry to the 1991 eruption confirmed to the north, floor the ASCT at 9° 37'. All of the above observations are consistent with the hypothesis that the eastern limb of the OSC experienced a volcanic event in the time frame between the 1989 ARGO survey and the *Alvin* dives in 1991. Further, it is proposed that the eastern limb of the OSC is actively propagating southward at the expense of the dying western limb, and that this propagation is due to more robust magmatic supply north of the OSC than exists

immediately south of it. Migration of this ridge-axis discontinuity is not unprecedented. Northward migration of the $9^{\circ} 37'$ discontinuity can be traced south to $\sim 9^{\circ} 20'$ and fault traces extend nearly thirty km off-axis suggesting that this discontinuity may have been in existence for nearly 500 thousand years [Macdonald, unpublished data]. That the $9^{\circ} 37'$ OSC now may be propagating in a southerly direction, opposite to its previous propagation direction is not unprecedented as a similar northward to southward change in propagation direction has been noted for the $11^{\circ} 45' \text{ N}$ OSC [Perram and Macdonald, 1990]. Considering the 400m offset distance of the $9^{\circ} 37'$ OSC, its apparent longevity and history of propagation, it is suggested that this discontinuity be classified as third-order.

3.5.4 Association of E-type MORB

Examination of geochemical data from MORB recovered in the wax core sample grid at $9^{\circ} 37' \text{ N}$ shows that the distribution of E-MORB is neither even nor symmetrical about the ridge-axis in this area (Figure 3-8). Perfit et al. [1994] similarly noted that distribution of E-MORB about the crestal plateau in the vicinity of $9^{\circ} 31' \text{ N}$ was not symmetrical about the ridge-axis, E-MORB being more abundant on the Cocos plate. Reynolds et al. [1992] also noted a similar asymmetric distribution of enriched MORB compositions in the 12° N region of the EPR. Geochemical data of samples recovered from the crestal plateau around the $9^{\circ} 37'$ OSC, however, show a very different distribution of E-MORB than noted previously by Perfit et al. [1994] or Reynolds et al. [1992]. Four times as many E-MORB were recovered on the Pacific plate (8) as were on recovered on the Cocos plate (2). Perhaps more intriguing is the observation that E-MORB on the Pacific plate are mostly concentrated within the latitudinal range of the

Figure 3-8: Map showing the distribution of rock core samples in the 9° 37' N region. Symbols are as in legend. Notice the asymmetric distribution of enriched MORB ($K_2O/TiO_2 > 10$) about the ridge-axis. Numbers above symbols indicate values of $K_2O/TiO_2 * 100$.

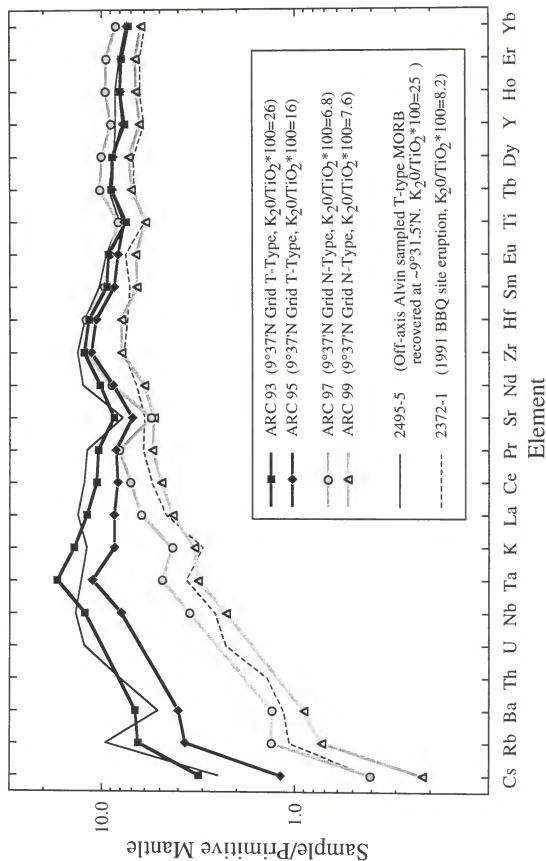


OSC, and occur closer to the ridge-axis (ASCT) than E-MORB sampled anywhere else on this second-order segment. Previous studies have noted the association of enriched MORB geochemistry with ridge-axis discontinuities on fast spreading MORs [Langmuir et al., 1986; Sinton et al., 1991], although it has been recently observed that this may not be the case in slower spreading environments [Batiza, 1996; and references therein].

Relative enrichments in incompatible elements between off-axis N-MORB and E-MORB recovered around the region of the 9° 37' OSC by wax core (Figure 3-9) suggest that they do not share a common parental magma. Instead, the E-MORB's greater enrichment of trace elements that are highly incompatible during mantle melting is more likely the result of N-MORB and E-MORB having distinct petrogenetic origins. The occurrence of dispersed, areally limited E-MORB has been frequently observed on the EPR [Langmuir et al., 1986; Sinton et al., 1991; Reynolds et al., 1992; Perfit et al., 1994]. E-MORB incompatible element abundances have been generally modeled to be the result of variable amounts of melt induced mixing of a heterogeneous, possibly veined, upper mantle source [Langmuir et al., 1986; Sinton et al., 1991; Reynolds et al., 1992; Perfit et al., 1994; Niu et al., 1996 and 1999]. Generally, E-MORB have been considered to result from lower extents of melting which, compared to greater extents of melting, preferentially samples lower melting temperature incompatible element enriched mantle heterogeneities.

EPR E-MORB are highly subordinate to N-MORB in terms of overall crustal volume, and have been observed to occur off-axis in areas where axial lavas are exclusively N-MORB [Reynolds et al., 1992; Perfit et al., 1994]. This observation led the authors to argue that preservation of E-MORB characteristics in erupted lavas

Figure 3-9: Primitive mantle-normalized spider diagram illustrating differences in the relative abundance of incompatible elements between N-MORB and E-MORB recovered from cretal plateau of segment D. Normalizing factors are after Sun and McDonough [1989]. Elements are listed from right to left in order of increasing incompatibility during mantle melting.



requires that the enriched melt avoid significant mixing with the N-MORB dominated AMC. To satisfy the above observations and the spatial association of E-MORB with axial discontinuities, Reynolds et al. [1992] proposed that E-MORB are always present in small volumes, but that they may only be erupted onto the seafloor under special conditions. These conditions include volcanic eruption either at structural offsets where magmatism is likely to be less robust, during the very early or late stages of a magmatic cycle when magmatic plumbing is poorly developed, or off-axis away from the influences of the AMC. Off-axis E-MORB have been identified around $9^{\circ} 30' \text{ N}$ [Perfit et al., 1994] and around $9^{\circ} 50' \text{ N}$ [Perfit et al., in prep], but E-MORB around the $9^{\circ} 37' \text{ OSC}$ occur much closer to the ridge-axis than in either of these two localities. In general, E-MORB distribution around the $9^{\circ} 37' \text{ OSC}$ are consistent with the conclusions of Reynolds et al. [1992] regarding E-MORB occurrence. It is proposed that E-MORB identified on the Pacific plate near the $9^{\circ} 37' \text{ OSC}$ relate to the northward propagation of the western OSC limb and subsequent magmatic waning of activity along that limb as the eastern limb advanced southward.

3.6. Conclusions

Submersible studies and wax core sampling in the region of a ridge-axis discontinuity at $9^{\circ} 37.1' \text{ N}$ on the EPR has led to the identification of a third-order OSC that marks a divide between axial magmatic and hydrothermal systems. These investigations make, for the first time, a direct link between magmatic activity and OSC propagation. Off-axis fault traces from northward propagation of this axial discontinuity suggest that it has been in existence for on the order of 10^5 years, further supporting its designation as a third-order discontinuity. Changes in lava and hydrothermal fluid

chemistry, ridge structure, volcanic morphology and apparent lava age all suggest that the eastern OSC limb and the ridge-axis north of it (Segments B and C of Haymon et al. [1991]) reflect a magmatic system that is currently more robust than that of the western OSC limb and ridge south of it (Segments D, E, and part of F). Geologic, chemical, biologic, and hydrologic data all suggest that a shallow intrusive event between 1989 and 1991, and likely a volcanic eruption, affected the eastern limb of the OSC. Chemical similarities between fresh sheet flows in the floor of the eastern limb's ASCT and the 1991 lava flow documented at 9° 46-51' further suggest that recent volcanic activity at 9° 37' may be related to the same magmatic pulse as the 9° 50' eruption.

A widening and southward extension of the ASCT on the eastern OSC limb by more than 400 m between 1989 and 1991 indicates that the eastern limb is actively propagating southward in response to this recent magmatic activity. The presence of extinct and brine venting sulfide structures in the ASCT of the western limb, and its generally older appearance further suggest that southward advancement of the eastern OSC limb may be at the expense of the western limb magmatic activity. Southward advancement of the eastern OSC limb further suggests that the fourth-order Segment "D" of Haymon et al. [1991] is likely to disappear as Segment "C" advances southward. It is further suggested that the DEVAL at 9° 35' which defines the southern terminus of segment D may be an artifact of its magmatic waning. The narrowness and overall character of the ASCT on the southern portion of Segment D suggests that despite the old appearance of lavas flooring it, the ASCT is not in the late stages of the proposed ASCT developmental cycle [Fornari et al., 1998b]. This suggests that magmatic plumbing along the western limb of the OSC may not have had the opportunity to fully develop into a

magmatically robust and mature system. Thus, Segment D may represent a transient fourth-order segment formed in response to propagation of a longer-lived axial discontinuity. Regional wax core sampling about the crestal plateau in the vicinity of the OSC has identified a concentration of E-MORB on the Pacific plate within ~1 km of the ridge-axis. The asymmetrical distribution of E-MORB in preference of the Pacific plate, their proximity to the OSC, and their unusual proximity to the ridge-axis (relative to other E-MORB recovered from the 9°-10° N segment) all suggest that E-MORB in this region are related to the western OSC limb. It is proposed that E-MORB are erupted in this region because the less robust magmatic system underneath the western OSC limb discourages efficient mixing and melt homogenization within the melt lens relative to regions with a robust, well developed and frequently replenished AMC.

CHAPTER 4 CONCLUSIONS

The detailed multi-disciplinary investigation of two recently active mid-ocean ridge segments in the eastern Pacific has resulted in a more detailed understanding of the nature of crustal accretion along these two ridge segments. The use of submersible and other deep water near-bottom systems, in conjunction with rock sampling by wax core, which provides better spatial control than traditional rock dredging, has resulted in a database of precisely located in situ geologic observations and samples. The combination of dense sampling, fine-scale spatial resolution of data, and temporal control introduced by the recent eruptive activity and longer term monitoring efforts make these investigations unique in MOR research. In particular, the addition of temporal control to the interpretation of MOR volcanic and hydrothermal activity on these two ridge segments has allowed for a more detailed account of the spatial and temporal associations between ridge structure, magmatic activity and geologic processes associated with oceanic crustal accretion within the neovolcanic zone of medium to fast spreading ridges.

Investigation of the CoAxial Segment on the Juan de Fuca Ridge, which is known to have erupted in 1993, showed that ridge segment has been volcanically active on at least three occasions between 1981 and 1993. Inter-flow chemical variations indicate that the three eruptive units do not have a direct genetic link with one another, and suggest the CoAxial Segment does not have a well-developed and connected magmatic plumbing system. Seismic monitoring of the 1993 eruptive event suggested that the lava

flow was emplaced after the intrusive dike had propagated laterally through the oceanic crust for 10's of kilometers, and inter-flow variations in lava chemistry suggest that lateral dike propagation may be a common emplacement mechanism on this segment. The structure of the segment is characteristic of a terrain where amagmatic extension plays a significant role in development of the ridge-axis. Volcanism within the neovolcanic zone of this segment does not appear to be particularly spatially focussed, with the three recent eruptions being emplaced over a kilometer-wide zone. Nonetheless, the eruption of some recent lava flows over preexisting, relatively fresh looking constructional volcanism suggests there is some structural control on the emplacement of lavas within the neovolcanic zone. The fissured character and poor focussing of axial volcanism are suggestive of a magmatically starved environment, yet there have been three eruptions in the last twelve years. This apparent contradiction may be indicative that magmatic activity on this medium spreading rate ridge segment is episodic, having short periods of relatively intense magmatic activity followed by longer periods of magmatic quiescence.

Extreme isotopic and elemental depletion of CoAxial segment lavas relative to other lavas from the Juan de Fuca ridge may indicate an origin for this ridge segment that is unique to the ridge. A regional survey of different morphological/structural/volcanic provinces within the central Juan de Fuca Ridge shows each to be fed by distinct mantle melt regimes. Generally, geochemical variation observed between the different melt regimes can be modeled by mixing of mantle source characteristics identified in regionally identified basaltic lavas. Correlation between long lived isotopes and trace element characteristics in central Juan de Fuca lavas suggest that at least some

component of regional upper mantle heterogeneity has antiquity to it, although trace element variability not correlated to isotopic ratios requires at least a third component of mantle heterogeneity. Close spatial association of ridge-axis lavas with distinct mantle geochemical signatures suggests that the spatial scale of heterogeneity may be quite small relative to the melt regimes identified.

Close spatial association of differently geochemically enriched/depleted lavas also occurs on the East Pacific Rise, but the relationship between ridge-axis structure and magmatic activity are very different than what is observed at the CoAxial Segment. Relative to the medium spreading, apparently magmatically starved CoAxial Segment, the fast spreading 9-10° N segment is magmatically robust, although spatial variability in melt supply to the ridge-axis does appear to exist. Data from a third-order overlapping spreading center (OSC) located at 9° 37' N suggest that there is a close association between magmatic activity and axial discontinuities, and further suggest that this OSC is a magmatic and hydrothermal divide between adjoining third-order ridge segments. The ridge-axis north of the OSC is currently in a robust phase of magmatic activity, while the ridge south of the OSC seems to be magmatically waning. North of the OSC, ridge-axis morphology is indicative of overall magmatic control, and south of the OSC structures related to amagmatic extension become more prominent. Temporal constraints that correlate a recent magmatic event to southward propagation of the eastern OSC limb establish, for the first time, a direct causal relationship between magmatic activity and evolution of ridge-axis discontinuities. The fourth-order segment associated with the OSC is likely a short lived, transitional feature that has not established a well-developed magmatic plumbing system. Support for this hypothesis includes a relatively poorly

developed ASCT floored with old lavas and an abundance of enriched MORB associated with the apparently dying western limb of the OSC. This data confirms that segmentation of melt delivery to the ridge-axis can be a significant factor in the evolution of ridge-axis structure along the fast spreading EPR.

APPENDIX FREQUENTLY USED ACRONYMS AND DESCRIPTIONS OF SAMPLING TECHNIQUES

Frequently Used Acronyms and abbreviations

MOR-	mid-ocean ridge
MORB-	mid-ocean ridge basalt
JdFR-	Juan de Fuca Ridge
EPR-	East Pacific Rise
OSC-	overlapping spreading center
ASCT-	axial summit collapse trough
ROV-	remotely operated vehicle
DEVAL-	deviation from axial linearity
REE-	rare earth elements

Sampling Techniques

Methods used to recover lava samples from the seafloor include *in situ* sampling by the manned submersible Alvin, remotely operated vehicle (ROV) and autonomous underwater vehicle (AUV). Precise location of sample acquisition in each of these cases is accomplished by means of a long-baseline transponder net which provides a precision accuracy to approximately 10 meters. *In situ* sampling by means of a deep submergence vehicle (submersible, ROV or AUV) provides the most accurate and informative means of sampling the deep seafloor. Other methods of lava sample recovery include a variety of surface deployed sampling techniques. These include wax coring (sometimes referred to as rock coring), rock dredging, and video guided grabs.

Wax coring: This technique utilizes a lead weight with a barrel affixed on its underside. This barrel is equipped with a "head" coated with wax. The apparatus is lowered to the seafloor at between 70 and 100 meters per minute, and upon contact with unsedimented

volcanic terrain pieces of volcanic glass become embedded in the wax. Typical recovery is between a few hundred milligrams and a few grams of volcanic glass. Precisional accuracy of sample location is dependent on depth, but typically is better than a few hundred meters.

Rock Dredging: Rock dredges lowered to the seafloor and dragged for variable distances thereby recovering rock fragments from lava flows on the seafloor. Sample recovery and precisional accuracy of sample recovery are highly variable and determined by surface ship location and setback calculations determined by the water depth and amount of cable spooled out.

Video guided grabs: This technique utilizes a hydraulic scoop with a real time video signal observed from a surface ship. This technique gives the precisional benefits of wax coring and excellent sample recovery. Additionally, the live video signal allows for *in situ* observations of the geologic associations of the sample recovered.

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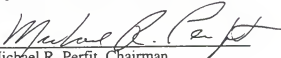
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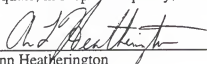
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Michael R. Perfit, Chairman
Professor of Geology

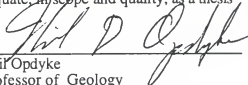
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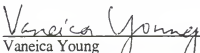
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